DOKUZ EYLÜL UNIVERSITY GRADUATE SCHOOL OF
NATURAL AND APPLIED SCIENCES

THE DEVELOPMENT OF A REGIONAL GROUNDWATER FLOW MODEL FOR THE TAHTALI WATERSHED

by
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İZMİR

THE DEVELOPMENT OF A REGIONAL GROUNDWATER FLOW MODEL FOR THE TAHTALI WATERSHED

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M.Sc. THESIS EXAMINATION RESULT FORM

We have read the thesis entitled **"THE DEVELOPMENT OF A REGIONAL GROUNDWATER FLOW MODEL FOR THE TAHTALI WATERSHED"** completed by **Deniz KARADAŞ** under supervision of **ASST.PROF.DR. ALPER ELÇİ** and we certify that in our opinion it is fully adequate, in scope and in quality, as a thesis for the degree of Master of Science.

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THE DEVELOPMENT OF A REGIONAL GROUNDWATER FLOW MODEL FOR THE TAHTALI WATERSHED

ABSTRACT

In this study, a regional groundwater flow model was developed for the Tahtalı watershed and surrounding area. The aim of the study was to determine the groundwater elevations of the region using a mathematical model, and to obtain the spatial distribution of the seasonal groundwater decline. Model was run as two-dimensional and steady-state conditions by application in unconfined aquifer.

Within the scope of this study, field work was conducted for the area between the Nif Mountain and the Tahtalı dam reservoir in May and October 2007. Water table depths were measured at certain monitoring wells and these measurements were put into a geographical information system along with other information relevant to the modeling study. Model inputs were obtained from field work results and from literature review. Hydraulic conductivity and groundwater recharge, which constitute the key model parameters, were considered as calibration parameters. The model domain was divided into four recharge, and seven hydraulic conductivity zones. A water budget-based precipitation-runoff model was used to calculate the recharge rate for the zone representing the Tahtalı watershed. Measured water table elevations of May were used as targets in the calibration of the model, whereas October measurements were used in the verification. Model boundary conditions were altered during the verification process. Also, recharge rates of the model were reduced based on precipitation records, considering the summer drought. Thereby, the model was run both for winter and summer conditions to calculate seasonal decline of the groundwater levels.

Groundwater elevation and flow direction maps were produced based on modeling results. Water budget results of the model revealed that groundwater recharge comprised 20% of the total water input for the entire study area. Furthermore, it can be noted that there is a hydraulic connection between the Nif

Mountain and the Tahtalı reservoir. The spatial distribution of the seasonal groundwater decline was limited. Decline values were lowest in the Cumaovası plain and were highest in areas with relatively lower hydraulic conductivities.

Keywords: Modeling, MODFLOW-2000, precipitation-runoff model, calibration, Izmir

ÖZ

TAHTALI HAVZASI İÇİN BÖLGESEL BİR YERALTI SUYU AKIM MODELİNİN GELİŞTİRİLMESİ

Bu çalışmada, Tahtalı havzası ve yakın civarı bölgeler için bölgesel bir yeraltı suyu akış modeli geliştirilmiştir. Çalışmadaki amaç; bölgedeki yeraltı suyu seviyelerinin matematiksel bir model yardımı ile belirlenmesi ve mevsimsel yeraltı suyu değişiminin alansal dağılımını belirlenmesidir. Model, iki boyutlu, kararlı akım koşullarında ve yüzeysel akifere uygulanarak çalıştırılmıştır.

Bu kapsamda, kuzeydoğuda Nif Dağ'ı ve güneyde Tahtalı gölü arasındaki çalışma alanında Mayıs 2007 ve Ekim 2007 dönemlerinde arazi çalışmaları yapılmıştır. Tespit edilen yeraltı suyu seviye ölçüm noktalarında su tablası derinlik ölçümleri yapılmış ve bu ölçümler model çalışmaları için gerekli olan diğer bilgiler ile birlikte bir coğrafi bilgi sisteminde düzenlenmiştir. Arazi çalışmalarından ve literatürden elde edilen bilgiler doğrultusunda, model girdileri belirlenmiştir. Modelin en önemli girdilerini teşkil eden hidrolik iletkenlik ve beslenim değerleri, kalibrasyon parametreleri olarak ele alınmıştır. Beslenim parametresi dört ve hidrolik iletkenlik yedi ayrı alt bölge için incelenmiştir. Tahtalı havzasını temsil eden beslenim alt bölgesinin yeraltı suyu besleniminin belirlenmesi için su bütçesi prensibine dayanan matematiksel bir yağış-akış modeli kullanılmıştır. Mayıs ayında ölçülen yeraltı suyu seviyeleri modelin kalibrasyonunda, Ekim ayı ölçümleri ise modelin doğrulanmasında kullanılmıştır. Doğrulama sürecinde modelin sınır koşulları değiştirilmiştir. Ayrıca yaz mevsiminin kuraklığını dikkate alarak beslenim değerleri de yağış verilerini baz alarak azaltılmıştır. Böylece hem kış, hem de yaz koşulları için model çalıştırılıp, mevsimsel yeraltı suyu düşümü hesaplanmıştır.

Yeraltı suyu akım model çıktıları ile yeraltı suyu seviye ve akım yönleri haritaları oluşturulmuştur. Modelin bütçe hesaplamalarına göre, beslenimin tüm çalışma alanına olan yeraltı suyu girdisinin %20 oluşturduğu tespit edilmiştir. Ayrıca, yeraltı suyu akım yönü haritasına dayanarak Tahtalı gölü ve Nif dağı arasında hidrolik bir bağlantının olabileceği bulunmuştur. Modelleme ile elde edilen mevsimel yeraltı suyu düşümleri incelendiğinde ise, düşümlerin konumsal olarak sınırlı değiştiği görülmüştür. En küçük değişimler Cumaovası'nda gözlenirken, en büyük düşümler hidrolik iletkenliği daha düşük olan bölgelerde gözlenmiştir.

Anahtar Kelimeler: Modelleme, MODFLOW–2000, yağış-akış modeli, kalibrasyon, İzmir

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CHAPTER ONE
INTRODUCTION

1.1 Study Objectives

The primary objective of this study was to develop and execute a numerical groundwater flow model for the semi-arid and environmentally stressed Tahtalı stream basin, which feeds the Tahtalı reservoir, an important water resource of the Izmir metropolitan area. The primary purpose of the model was to determine the seasonal decline of the groundwater table at the end of the dry summer season. Also, the groundwater flux to the neighboring Torbalı sub-basin was estimated using the model. The developed model can also be used for a subsequent watershed contaminant transport modeling study, where the output of the groundwater flow model is used as input to the transport model. A groundwater flow model of the area may also prove useful for the evaluation of climate change impact scenarios. It is therefore an essential precursor for any kind of hydrological study.

1.2 Scope of the Study

A regional groundwater flow model was developed for the Tahtalı stream basin and surrounding areas. The study comprised of two main tasks; the first task was to collect necessary data, both from existing databases and previous studies, and during field trips. The second task was the theoretical and practical development of the groundwater flow model. Field trips were planned to explore the study area and determine possible groundwater monitoring locations and to subsequently measure groundwater levels. A geographic information system (GIS) was utilized for organizing and processing of various data and for map production and illustration of modeling results. After the numerical groundwater flow model was set up, the model was calibrated and also verified using two different sets of observed groundwater level data. Calibration statistics and other indicators of model performance were calculated for both the calibrated and verified models to objectively assess the validity of the groundwater flow model. Contour maps of hydraulic head and maps showing groundwater flow directions were generated and interpreted with respect to local hydrogeology, groundwater withdrawals and the groundwater flux within the basin. Finally, the spatial distribution of seasonal groundwater decline was determined using the developed model.

CHAPTER TWO
LITERATURE REVIEW

Water is the source of all life on earth. The distribution of water, however, is quite variable; many locations have plenty of it while others have very little. Water exists on earth as a solid (ice), water vapor, and in liquid form above and below the ground surface. Both surface and subsurface waters originate from precipitation, which includes all forms of moisture from clouds, including rain and snow. Part of the precipitation water runs off over the land (surface runoff), infiltrates and flows through the subsurface (subsurface flow), and eventually finds its way back to the atmosphere through evaporation from lakes, rivers, and the oceans; transpiration from trees and plants; or evapotranspiration from vegetation.

Not all subsurface water is groundwater. Groundwater is all the water that has penetrated the earth's surface and is found in one of two soil layers. The one nearest the surface is the vadose zone, where gaps between soil particles are filled with both air and water. Below this layer is the saturated zone, where the gaps are filled with water. The water table is the boundary between these two layers. As the amount of groundwater water increases or decreases, the water table rises or falls accordingly. When the entire area below the ground is saturated, flooding occurs because all subsequent precipitation is forced to remain on the surface.

Groundwater supplies wells and springs, and it replenishes streams, rivers, lakes and also provides fresh water for irrigation, industry, and communities. It has advantages and disadvantages when comparing with surface water. The advantages of using groundwater can be listed as follows:

1. Significantly better quality compared to surface water and little to no water treatment costs.
2. Passage through soil and granular materials allows the filtering of microorganisms and minute particles, as well as the attachment of organic compounds and some metals to clay minerals.
3. Temperature and chemical quality are relatively constant over time.

4. Dispersion of pollution is slower.
5. Sediment content is generally negligible.
6. Supply is generally unaffected by short-term fluctuations in climate.

The disadvantages of groundwater:

1. Dissolved mineral content and hardness are higher than surface water.
2. Management is more difficult.
3. Exploration and characterization of groundwater resources require advanced skills and methods.
4. Once groundwater is contaminated, subsurface cleanup is difficult and costly, and the application of cost-ineffective pump-and-treat methods may be the only viable option.

2.1. Modeling of Groundwater Flow

Groundwater modeling can be defined as the quantification and simulation of the natural movement of groundwater through any porous media. This can be achieved by physical or mathematical means. Modeling plays an extremely important role in the management of water resources. Groundwater models, which replicate the groundwater flow process at the site of interest, can be used to complement monitoring studies in evaluating and forecasting groundwater flow and transport. However, every reliable groundwater model is based on accurate field data and decent prior knowledge of the site. The groundwater modeling process is summarized in Figure 1.1.

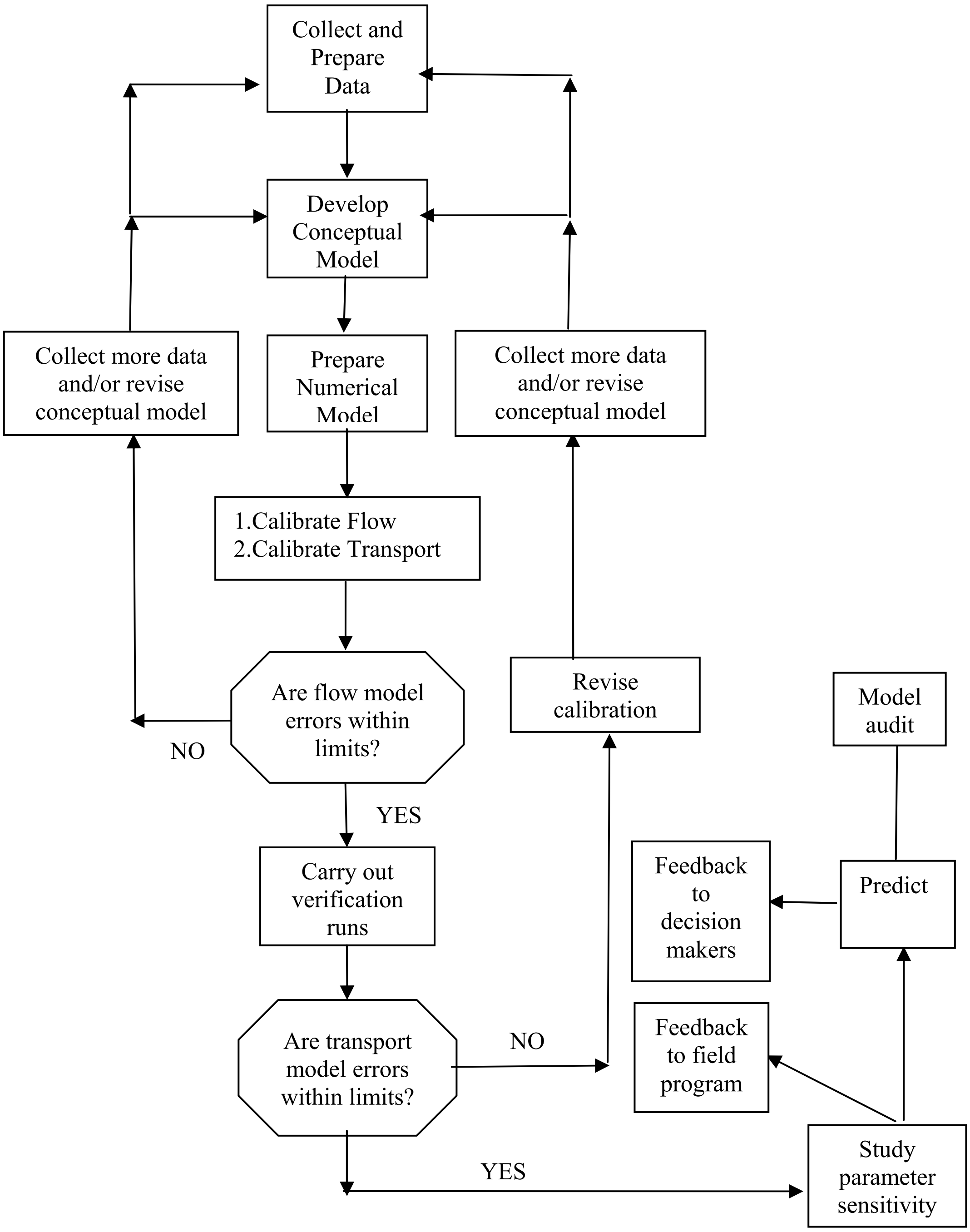

Figure 1.1 The groundwater modeling process

2.1.1 Types of Groundwater Models

There are several ways to classify groundwater flow models. Models can be either transient or steady-state, confined or unconfined, and consider one, two or three spatial dimensions. In setting up the grid of a numerical model, the classification that is most relevant is one based on spatial dimension (Anderson, 1990). In general, there are three types of models to be used for modeling as physical, mathematical, empirical methods. A mathematical model was used in the study presented here.

A mathematical model is an exact or approximate solution to the governing equations of the process. Mathematical models of groundwater flow, which are also called white box model, have been in use since the late 1800s. Fundamental theories, principles and some simplifying assumptions are used to derive equations. Simplifying assumptions must always be made in order to construct a model because the field situations are too complicated to be simulated exactly. Usually the assumptions necessary to solve a mathematical model analytically are fairly restrictive. For example, many analytical solutions require the subsurface medium to be homogenous and isotropic. To deal with more realistic situations, it is usually necessary to solve mathematical model approximately using numerical techniques.

The general governing equation for three-dimesional, transient groundwater flow in a heterogeneous and anisotropic aquifer is given as:

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial z}\right)=S_s \frac{\partial h}{\partial t} \qquad \text{(Eq. 1.1)}$$

Here h represent the hydraulic head; x,y,z and t represent the spatial dimensions and time, respectively; K_x, K_y, K_z are the hydraulic conductivities in the x,y and z directions and S_s is the specific storage of the aquifer. The derivation of this equation is based on the application of the mass balance principle on a finite element representing the saturated porous medium and the substitution of groundwater flux terms with Darcy's law.

Mathematical models can be grouped into of analytical and numerical models:

2.1.1.1 Analytical Models

Analytical models are exact solutions to the differential equations written in terms of elementary or known functions. For example, the governing equation can be written like below for one-dimensional, transient groundwater flow in a homogeneous, confined aquifer:

$$S_s \frac{\partial h}{dt} = K \ \frac{\partial^2 h}{\partial x^2} \qquad \text{(Eq. 1.2)}$$

Here, initial and boundary conditions need to be defined, and a function of hydraulic head, which depends on space and time is obtained. On the other hand, the "Theis solution" is a well-known analytical model and is widely used. The Theis solution is formulated as:

$$S(u) = \frac{Q}{4\pi T} W(u) \qquad \text{(Eq. 1.3)}$$

In this equation, W represents the "well function", S is the drawdown and T is the transmissivity of the aquifer. Analytical models provide continuous solutions over the model domain. Analytical models have some advantages and disadvantages over numerical models. One of the advantages is that these models are computationally very efficient and provide accurate solutions. Also, they are appropriate for limited data and useful for quick initial estimation of systems behavior. However, if the model geometry is complicated, analytical models are difficult to apply. Also, the analytical solution of the governing equations may require sophisticated math techniques. Analytical models often have many limitations, leading to over-simplified solutions, and they are usually restricted to 1-D or 2-D.

2.1.1.2 Numerical Models

Numerical models allow analysis of flow or transport solutions, if the complexity of the mathematical model prevents an analytical solution. Numerical modeling techniques are used to solve large set of equations, which describe the physical flow processes in an aquifer. There are two numerical techniques of numerical models which are called finite differences and finite elements methods (Figure 1.2). These two approximate methods provide a rationale for operating on the differential equations that make up a model and for transforming them into a set of algebraic equations.

Numerical modeling provides a discrete solution over the model domain used by algebraic equations. It uses iterative methods or direct methods for the approximate solution. For many problems numerical solution is more realistic than the analytical solution. In this case, generally numerical models are preferred to use in mathematical model. Values are calculated at only a few points by the numerical models.

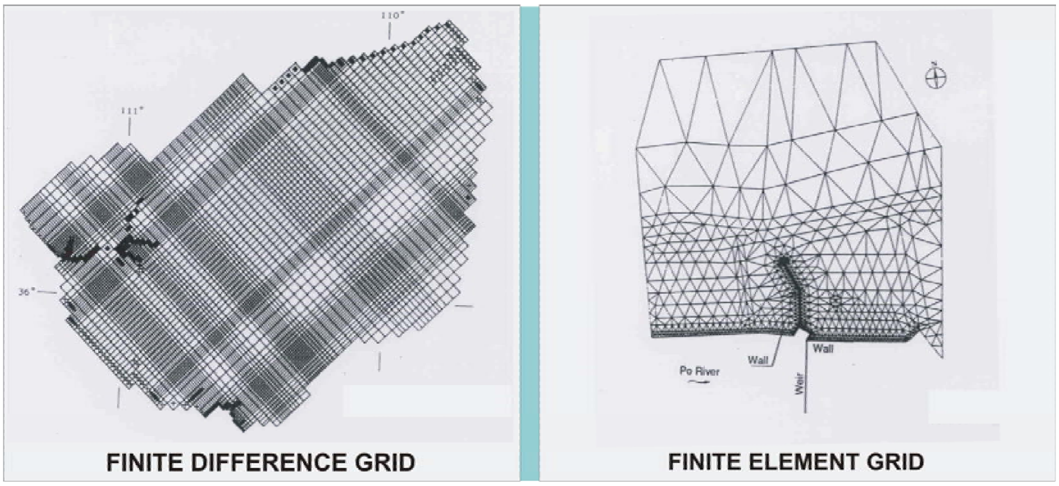

Figure 1.2 Types of numerical models (Kumar, 2005)

Numerical approximations to the governing differential equations are solved numerically over a grid. The choice between a finite difference and finite element model is usually a matter of preference and depends sometimes on the problem to be solved. The numerical solution methods used:

2.1.1.3 Finite Difference Method

The Finite-Difference Method is one of the oldest methods for solving partial differential equations. The computational domain is discretized by rectangular or quadrilateral cells (Figure 1.3). Often, the cell dimensions Δx and Δz are constant or even $\Delta x = \Delta z$. The unknown defined in nodes, which are placed at centers of the cells or at the intersection points of cell boundaries (Hinkelmann, 2008). Depending on the finite-difference model, groundwater heads or concentrations are calculated as discrete values at the grid nodes, or at the center points of cells. (Spitz & Moreno, 1996)

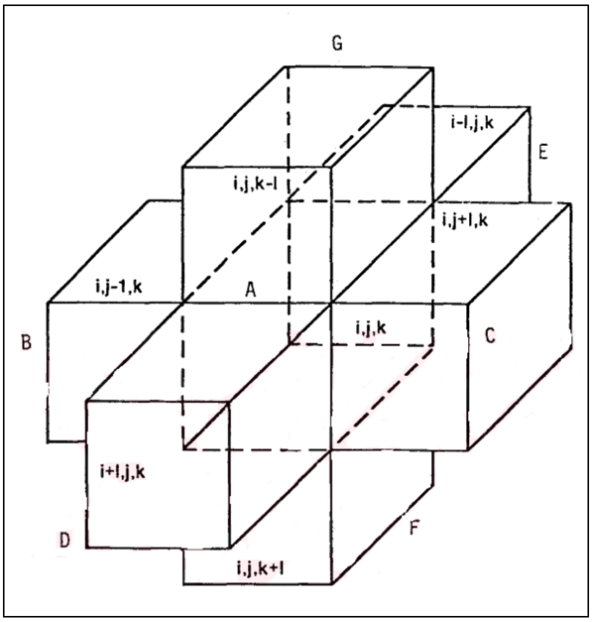

Figure 1.3 Illustration of a finite-difference computational molecule

From the geometrical point of view, it is obvious that complex boundaries or complex inner structures can only be reproduced in a very simplified way by step functions. Derivatives of the unknown function "e" can be developed with the help of a Taylor-series expansion, shown here for the *x* direction. For simplicity's sake, constant Δx=Δz are assumed.

$$e_{i+1j} = e_{i,j} + \Delta x \frac{\partial e_{i,j}}{\partial x} + \frac{(\Delta x)^2}{2} * \frac{\partial^2 e_{i,j}}{\partial x^2} + \frac{(\Delta x)^3}{6} * \frac{\partial^3 e_{i,j}}{\partial x^3} + \frac{(\Delta x)^4}{24} * \frac{\partial^4 e_{i,j}}{\partial x^4} + 0(\Delta x^5) \quad \text{(Eq. 1.4)}$$

In the following, the principle use of the finite-difference method is explained using flow process in groundwater. Furthermore, one dimensional problem is considered with a constant hydraulic conductivity K_f and a constant storage term S_0 and without sink and source terms.

$$S_0 \frac{\partial h}{dt} - div - (K_f gradh) = 0 \quad \Longleftrightarrow \quad \frac{\partial h}{dt} - \frac{K_f}{S_0} \frac{\partial^2 h}{\partial^2 x^2} = 0 \qquad \text{(Eq. 1.5)}$$

The method is established the continuity equation for each cell taking into account initial and / or boundary conditions. Depending on the governing equation, inflow and outflow are calculated for each cell. After expressing the continuity equation for unknown heads, set of the equation is solved for each cell. Numerous groundwater flow codes exist that are can solve the general equations of groundwater flow using the finite-difference methods. One of them is MODFLOW-2000 (Harbaugh, 2000).

MODFLOW-2000 is the third major release of the popular U.S. Geological Survey 3-D finite difference groundwater flow model. MODFLOW was originally programmed under the FORTRAN-77 language environment to solve the finite-difference equations that represent 3-D saturated groundwater flow. It was first developed by McDonald and Harbaugh (1984) of the U.S. Geological Survey in 1984 and was updated four times resulting with the versions MODFLOW-88, MODFLOW-96, MODFLOW-2000 and MODFLOW-2005. At the same time many new packages were added into the code, which can simulate the hydrologic problems much better than ever. These packages can be used separately by main program during calculating the model and each package is divided into different modules and each module executes different procedure to finish certain part of simulation such as defining model, allocating memory, reading data, formulating equations (Wang et.al. 2007).

MODFLOW has a modular structure that allows it to be modified to adapt the code for special applications. It simulates steady and transient flow in an irregularly shaped flow system in which aquifer layers can be confined, unconfined, or a combination of confined and unconfined. Flow from external stresses, such as flow to wells, aerial recharge, evapotranspiration, flow to drains, and flow through river beds, can be simulated. Specified head and specified flux boundaries can be simulated as can a head dependent flux across the model's outer boundary that allows water to be supplied to a boundary block in the modeled area at a rate proportional to the current head difference between a source of water outside the modeled area and the boundary block. In addition to simulating ground-water flow it incorporates related capabilities such as solute transport and parameter estimation. The groundwater flow equation is solved using the finite-difference approximation. The flow region is subdivided into blocks in which the medium properties are assumed to be uniform. (USGS, 2008)

2.1.1.4 Finite Element Method

The application of the finite element (FE) method to groundwater problems is a relatively recent development with respect to the finite difference method. The finite element method is implemented with a variety of element types, but the triangular element is good beginning point for describing the method. (Anderson & Wang, 1990)

The FE model differs from the FD model by approximating the flow equation by integration rather than differentiation. As in the FD model, the model area is subdivided into sub-areas, called elements. One normally chooses triangular elements as sub-areas. Since there are basically no restrictions on the shapes of the elements, the model user is more flexible in the model discretization than when using the finite difference scheme. (Spitz & Moreno, 1996)

2.1.2 Data Requirements for Groundwater Modeling

Compiling the field data relevant to the assembly of the groundwater flow model is a significant step in modeling. Data requirements for groundwater modeling can be classified in two sections; the physical and hydrologic framework. The first step of a model study consists of collection and evaluating relevant data on flow system under investigation. Input data for the model are used for (Spitz & Moreno, 1996):

1. Problem definition (material properties and geometry of hydraulic units)
2. Numerical requirements (initial conditions, boundary conditions and transient conditions)

3. Modeling requirements (calibration, validation, and definition of alternate scenarios)

Data in the physical framework define the geometry of the system including thickness and real extent of each hydrostratigraphic unit. Data within the hydrologic framework include information on heads and fluxes, which are needed to formulate the conceptual model and check model calibration. Hydrogeologic data also define aquifer properties and hydrologic stresses. They include pumping, recharge and evapotranspiration. Recharge is the one of the most difficult parameters to estimate. (Anderson & Woessner, 1990)

The physical framework consists of all geological information about the natural system such as a geological map showing cross-sections, vertical profiles, fault lines and formations, a topographic map showing surface water bodies, residential and industrial areas, surface elevation contours, etc., contour maps showing the elevation of the base of aquifers and confining beds, isopach maps showing the thickness of streams and lake sediments. The physical framework basically defines the geometry of the system including the thickness of the hydrostratigraphic units.

Data on hydraulic heads, fluxes, precipitation, evapotransipiration are included within the hydrological framework. Hydrological data also define hydrologic stresses such as pumping, recharge and evapotranspiration. Hydrological data can come in the form of water table and potentiometric maps for the aquifers of interest, hydrographs of groundwater head and surface water levels and discharge rates, maps showing hydraulic conductivity and transmissivity distribution, spatial and temporal distribution of rates of groundwater recharge, groundwater pumping, natural groundwater discharge and evapotranspiration.

2.1.3 Model Calibration and Verification

2.1.3.1 Model Calibration

Model calibration is defined as systematically changing values of model input parameters in an attempt to match field conditions within some acceptable criteria.

Calibration is accomplished by finding set of parameters, boundary conditions, and stresses that produce simulated heads and fluxes that match field-measured values within a pre-established range of error. Finding set of values amounts to solving what is known as the inverse problem. In an inverse problem the objective is to determine values of parameters and hydrologic stresses from information about heads, whereas in forward problem system parameters such as recharge rate specified and the model calculates heads.

The objective of the calibration is to minimize this error, sometimes called the calibration criterion. Calibration statistics can be expressed in many ways but the most common are listed below:

1. Mean error (ME) is the arithmetic mean of differences between measured and simulated heads (residuals) where is the number of calibration values. Caution should be exercised when interpreting this error as negative and positive residuals may cancel out and yield a low error.

$$ME = \frac{1}{n}\sum_{i=1}^{n}(h_m - h_s)_i \qquad \text{(Eq.1.6)}$$

   In this formulation, h_s indicates simulated heads and h_m is measured heads. The ME is simple to calculate but is usually not a wise choice because both negative and positive differences are incorporated in the mean and may cancel out the error.

2. Mean absolute error (MAE): the arithmetic mean of the absolute value of the differences in measured and simulated heads.

$$MAE = \frac{1}{n}\sum_{i=1}^{n}\left|h_s - h_m\right| = \frac{1}{n}\sum_{i=1}^{n}\left|ei\right| \qquad \text{(Eq. 1.7)}$$

   As the name suggests, the mean absolute error is an average of the absolute errors $e_i = h_s - h_m$, where h_s is the calculated and h_m the measured value.

3. Root mean squared (RMS) error: the standard deviation of the differences in measured and simulated heads. The RMS is usually thought to be the best measure of error, if the errors are normally distributed. The maximum acceptable value of the calibration criterion depends on the

magnitude of the change in heads over the problem domain. If the ratio of the RMS error to the total head range in the system is small, the errors are only a small part of the overall model response.

$$x_{rms} = \sqrt{\frac{1}{n}\sum_{i=1}^{n} x^2{}_i} = \sqrt{\frac{x_1^2 + x_2^2 + \ldots\ldots + x_n^2}{n}} \qquad \text{(Eq.1.8)}$$

The RMS is usually thought to be the best measure of error if errors are normally distributed.

The three measures of error discussed above quantify the average error in the calibration but say nothing about the distribution of error. For instance, comparison of head contours gives a purely qualitative and subjective indication of spatial distribution error. A quantitative analysis of the distribution or error should be part of calibration assessment. The error in the residuals should be randomly distributed over the grid or contours (Anderson & Woessner, 1990).

There are basically two ways to estimate model parameters and solve the inverse problem: the manual trial-error adjustment of parameter and automated parameter estimation. In trial and error calibration, parameter values initially assigned to each node or element in the grid are adjusted in sequential model runs to compare simulated heads or flows to the calibration target. On the other hand, the automated calibration method is performed using specially developed codes that use either a direct or indirect approach to solve the inverse problem. In a direct solution, the unknown parameters are treated as dependent variables. This means that values for head must be input for all nodes. Heads are known only at points where there are observation wells, making it necessary to estimate heads elsewhere in the grid, usually by interpolation methods like kriging. The solution minimizes the nodal mass balance error caused by using these heads and the model parameter values. The indirect approach is similar to performing trial-error calibrations in that the forward problem is solved repeatedly in an automated fashion.

A good example for an automatic calibration code is PEST, which is a calibration tool, developed by John Doherty of Watermark Computing that works with all types of models that use one or more input files and produce one or more output files (Doherty, 2004). PEST works by using a template file that is a copy of the MODFLOW file containing parameters to be estimated. The parameters are replaced by a special code that tells PEST where to get the parameters. This means that the parameters to be estimated must be the ones written to the MODFLOW file (usually the BCF Package file or boundary condition files).

2.1.3.2 Model Verification

Model verification is a test of whether the model can be used as a predictive tool, by demonstrating that the calibrated model is an adequate representation of the physical system. Owing to uncertainties in the model input data, the set of parameter values obtained after the calibration process may not accurately represent actual field values. Consequently, the calibrated parameters may not accurately represent the system under a different set of boundary conditions or hydrologic stresses or the calibrated solution may be non-unique.

Model verification helps establishing greater confidence in the calibrated model. In a typical verification exercise, values of parameters and hydrologic stresses determined during calibration are used to simulate a transient response for which an independent and different set of field data exists. If the calibrated parameters were changed significantly during verification, it may not be possible to match the calibration targets using the new parameter values. In this case it will be necessary to repeat the process until a set of parameter values is identified that producers a good match to both the calibration and what were intended to be verification targets. If it is necessary to adjust parameters during verification, the verification becomes a second calibration and another independent data set is needed to perform the verification. Verification is accomplished when the verification targets match without changing the calibrated parameter values. (Anderson & Woessner, 1991)

2.2 Synopsis of Literature about Previous Regional Groundwater Modeling Studies

Atilla (1998) developed a transient groundwater flow model for the confined aquifer under the Afyon Plain in Turkey. The spatial and temporal extent of hydraulic head over the plain was simulated using MODFLOW. According to the piezometric level decline and water quality degradation conditions, the prediction of the consequences of the overexploitation requires the identification of the current head distribution. The hydraulic head distribution declines from NW to SE over the plain. The model shows that there is an increase in the decline of the piezometric levels after the year 1976 when intensive groundwater exploitation is started, and after 1990 when the exploitation is considerably increased. It is simulated that the hydraulic head is decreased 5 to 10 m in some parts of the plain front the year 1965 to 1998. Under these conditions, groundwater usage In the Plain should he regulated to establish the natural hydraulic balance and, the termination of uncontrolled ground water exploitation.

Ayenew, Demlie & Wohnlich (2007) conducted a numerical modeling study for the groundwater system in the Akaki catchment of central Ethiopia. A 3-D steady-state finite-difference groundwater flow model was developed to quantify the groundwater fluxes and analyze the subsurface hydrodynamics in the Akaki catchment by giving particular emphasis to the well field that supplies water to the city of Addis Ababa. The model was calibrated using head observations from 131 wells. The simulation was made in a two layer unconfined aquifer with spatially variable recharge and hydraulic conductivities under well-defined boundary condition. The result indicated that the groundwater flows regionally to the south converging to the major well field.

Juckem, Hunt & Anderson (2006) provided extensive data that scale effects of hydrostratigraphy and recharge zone on base flow. This study's objective was to present a methodology for estimating a critical basin size, above which base flows appear to be relatively less sensitive to the spatial distribution of recharge and hydraulic conductivity. Influence of recharge zonation and hydrostratigraphic layering on base flow was determined using MODFLOW for the Coon Creek Watershed, which is located in the Wisconsin, USA. This model was set up as three-dimensional and for steady-state conditions. The results showed that there is a scale effect that influences the relative importance of recharge and hydraulic conductivity such that at some scale, the influence of spatial parameter variability on base flow diminishes and can be approximated using a simplified representation.

Elçi, Gündüz & Şimşek (2007) developed a mathematical flow model for the water table aquifer of the Torbalı plain in Izmir. This two-dimensional model was set up for steady-state conditions, and was executed using MODFLOW-2000. Groundwater levels were measured in the study area at 28 monitoring points. Hydraulic conductivity and aquifer recharge rates were used for model calibration. Groundwater flow directions and the water budget for the Torbalı Basin were determined by this modeling study. According to the modeling results, the plain receives groundwater influx from the limestone units in the south and the Gurgur Mountain in the east of Torbali, in addition to surficial recharge originating from precipitation.

He, Takase & Wang (2007) used a three-dimensional finite element model, this study characterizes groundwater flow in a costal plain of the Seto Inland Sea, Japan. The model characterization involved taking field data describing the aquifer system and translating this information into input variables that the model code uses to solve governing equations of flow. Geological geometry and the number of aquifers have been analyzed based on a large amount of geological, hydrogeological and topographical data. Results of study demonstrate a high correlation between the ground surface elevation and the groundwater level in the shallow coastal aquifer. For calibrating the numerical groundwater model, the groundwater flow was simulated in steady state. In addition, the groundwater level and trend in the transient state has also been elucidated. The numerical result provides excellent visual representations of groundwater flow, presenting resource managers and decision makers with a clear understanding of the nature of the types of groundwater flow pathways. Results build a base for further analysis under different future scenarios.

McAda & Barroll (2002) developed a three dimensional, finite difference groundwater flow model for the Middle Rio Grande basin in New Mexico, U.S.A. The purpose of the model was to integrate the components of the groundwater flow system, including the hydrologic interaction between the surface water systems in the basin, to better understand the hydrogeology of the basin and to provide a tool to help water managers plan for and administer the use of basin water resources. Groundwater flow in the Middle Rio Grande basin was simulated from 1900 to March 2000. Steady-state conditions were assumed to exist prior to 1900, which was used as initial conditions for the transient simulation period of 1900-2000. The model was calibrated using a judgmental trial-and-error procedure of adjusting aquifer properties and boundary conditions in an effort to minimize the difference between measured and simulated water-level data and flow data by MODFLOW-2000. Also recharge parameters were defined as different kind of types such as mountain-front recharge, tributary recharge and subsurface recharge. In additionally, hydraulic conductivity definitions were classified horizontal and vertical along model columns and rows. Other parameter of model such as, specific storage which was estimated to be 2×10^{-6} per foot in the model and specific yield was estimated to be 0.2.

Moustadraf, Razack & Sinan (2008) developed a numerical and transient model which related to intensive pumping during the periods of drought, which was forced abandonment of wells due to the seawater intrusion in the aquifer of the Chaouia Coast of Morocco. With respect to climatic fluctuations, precipitation and temperature data were analyzed. Before modeling, conceptual model area was constructed as the top of the layer was represented by the topographic surface. It constitutes the recharge area by precipitation to the system. The bottom layer corresponds to the Paleozoic bedrock which is represented by a no-flow boundary. Hydraulic conductivity and recharge was used for calibration. Hydraulic conductivity was estimated by interpolation. The steady-state simulation is based on the lower and higher groundwater level periods in 1971. The aim of this simulation is to calibrate the model by adjusting the spatial distribution of the hydraulic conductivity and of the recharge. The transient simulation, based on the calibration obtained in steady-state simulation, aims at simulating the evolution vs. time of the groundwater flow of the aquifer. The numerical modeling showed that the severe degradation of the resource was primarily related to intensive pumping which was 7meters during periods of drought. This pumping has induced seawater intrusion into the aquifer and consequently the abandonment of wells contaminated by salt water.

Palma & Bentley (2007) constituted a regional scale groundwater flow model which was simulated using transient and steady-state numerical models for the Leon-Chinandega aquifer in Nicaragua used by Visual MDOFLOW. The study focused on a quantitative assessment of the potential of the aquifer as a source of water for irrigation. The purpose of this work was to study the groundwater flow system in a sub-basin of the Leon-Chinandega aquifer using transient and steady-state numerical groundwater- flow models and to investigate the effects of further groundwater development. This model was calibrated by model discharge and transmissivity. The transient simulations were run for 10 years, using the results from either 1970–1971 or 2004–2005 to remove the influence of the steady-state simulation initial condition. Two different flow systems are identified in the Leon-Chinandega aquifer. The first one was a deep system, recharged in the cordillera and then discharged in the central and lower plain, either as base flow or to pumping wells. The second was a shallow local flow system, recharged in the central and lower plain which was discharged into the rivers or pumping wells. Simulations indicated that groundwater from deep wells is recharged at high elevations, corresponding to the deep flow system. Shallow wells mostly capture groundwater that was recharged locally, but there was also an indication that mixing of the regional and local system can occur.

Sakiyan & Yazicigil (2004) studied the aquifer system of the Küçük Menderes basin for sustainable development and management of an aquifer system. The spatial distributions of the hydrogeological parameters and recharge were estimated by geostatistical methods and hydrologic simulations. A finite-difference groundwater flow model was used to represent the unconfined flow in the aquifer system. This model was calibrated in sequential stages as a steady-state followed by transient condition. The study’s objective was to develop a groundwater management plan using the groundwater flow model. Alternative groundwater management scenarios were developed to determine the safe yield for the Küçük Menderes aquifer system.

Soyaslan (2004) generated a groundwater flow modeling map of Yalvaç Basin based on three dimensional, steady state condition, and finite difference methods by MODFLOW. A numerical groundwater flow model of the Yalvaç basin, which is a closed basin, was developed to determine the amount of groundwater discharge to the Eğirdir Lake. The basin has been modeled as different four layer aquifer system. In the bottom, they identified that there is semi-confined karstic aquifer and less permeability and storage capacity. In this study, a numerical groundwater flow model of the Yalvaç basin, which is a closed basin, was created to determine amount of groundwater discharge to Eğirdir Lake and to determine the aquifer parameters according to the validated conceptual hydro geological model of the study area. Therefore this model has been calibrated by average groundwater level observations, spring discharges and drain level of 2000. As a result discharge amount has been determined as yearly total of 114×10^6 m^3 to Eğirdir Lake.

The El Paso Water Utilities Public Service Board (2002) in the U.S.A. prepared a report about groundwater modeling study results for the Cãnutillo Wellfield. The purpose of this model was to provide insight into the groundwater system of the Mesilla Bolson and as such provide information to be used in water resources planning. The grid of the model domain was made uniform at a spacing of approximately 200 meters. Additional canals, drains, and laterals were added. During development of a groundwater flow model, parameters such as hydraulic conductivity were input to the model based on available test data, knowledge of the aquifer hydrogeology, and interpolation between known values. On the other hand, based on the groundwater flow model, a contaminant transport model was developed to provide more reliable estimates of changes in water quality over time than can be produced analytically. The transport model covered the same area as the Cañutillo flow model and used the solved head distribution from the flow model as an input. Calibration of the groundwater transport model required that concentrations in individual wells matched over time through changing selected parameters, boundary conditions, and initial concentrations. A baseline simulation was completed based on the best available information on model parameters and starting conditions. This simulation was then compared to subsequent simulations to gage model improvement with parameter changes. Model parameters such as porosity, mountain- and slope-front recharge concentration, and irrigation recharge concentration were altered until the best solution was achieved.

The Miami-Dade and Sewer Department of Environmental Resource Management (2001) prepared a report, which was about risk assessment and groundwater modeling of the Miami-Dade area in the U.S.A. The Miami-Dade well field consisting of fifteen water supply wells had a maximum daily permitted allocation of 587.45×10^3 m^3/day. The two WTPs have a combined permitted capacity 852.75×10^3 m^3/day and utilize conventional lime softening treatment, followed by filtration and disinfection. While this treatment was adequate for groundwater sources, it would not be sufficiently protective if the source were under the direct influence of a surface water body. Although the well field and WTP are currently limited by permit to 587.45 and 852.75×10^3 m^3/day, respectively, MDWASD (The Northwest Wellfield was Miami-Dade Water and Sewer Department) indicated that the planned future capacity of the North Well Field is 890.65×10^3 m^3/day. Thus, this value would be used in all future analyses within this report. The numerical model had to be used to more precisely estimate the travel times and paths of groundwater. Purpose of using the groundwater model was to estimate modeling of Cryptosporidium provides a conservative and protective approach the 180-day and 230-day particle travel time distances in the vicinity of the NWWF. The pumpage from the NWWF was simulated at MDWASD's planned future withdrawal of 890.65×10^3 m3/day. The model results were used to develop travel-time contour plots based on particle tracking simulations using the MODFLOW post-processor MODPATH.

Encon (2005) prepared a comprehensive environmental impact assessment report for a planned gold mine located in Efemçukuru, Izmir. Groundwater resources, groundwater levels, groundwater quality and flow direction around the area was searched and reported. Groundwater resources were classified as three sources such as wells, creeks and drilling wells. Groundwater level was measured due to seasonal alteration. Groundwater flow and hydraulic gradient were determined by the drilling and alluvial wells which were measured water levels. Hydraulic conductivity was obtained with some tests. While groundwater flow was being identified, seasonal drawdown of water levels was considered. According to the results, the groundwater flow increased to the eastern of the mine. Also hydraulic gradient increased due to topography. The contamination would be transported towards to the Torbalı Pain.

Weiss & Gvirtzman (2007) studied 20 to 30 years of precipitation and spring discharge records to reconstruct the transient character of yearly recharge using a groundwater flow model. Four different sites within the Yargon-Taninim aquifer, which is the most important resources of fresh water of Israel, were chosen for building conceptual and numerical hydrogeological models. Transient, finite difference numerical groundwater flow models were developed for four separate perched karstic aquifers in the Judean Samarian Mountains in Israel using MODFLOW-2000. The resulting numerical ground water flow model was calibrated to both the rainfall data (using precipitation- recharge relationships) and the spring discharge data. Precipitation-recharge functions were estimated by numerical modeling. Best fitting between measured and computed spring hydrograph data allowed to develop a set of empirical functions relating measured precipitation to recharge to the aquifer.

CHAPTER THREE
DESCRIPTION OF THE STUDY AREA

3.1 General Description

The groundwater flow model was developed for the Tahtalı stream basin and the urban area between the basin and the Bay of Izmir. The general location map of the basin and the study area is given in Figure 3.1. The Tahtalı dam reservoir is one of the most important drinking water resources for Izmir. It meets about 36% of Izmir's water demand and is located 40 km south of Izmir and 5 km east of the town of Gümüldür (38°08' N; 27°06' E). The basin has a drainage area of 550 km^2 and is surrounded by the Sandallı Mountain in the west and the Nif Mountain in the northeast. A total of 44 ephemeral creeks, streams and their tributaries drain through the basin and feed the Tahtalı reservoir. The major inflows to the reservoir are from the north via Tahtalı and Şaşal streams. The Şaşal stream contributes 25% whereas the Tahtalı stream contributes 75% to the total inflow. The discharges of the other four inflowing streams are negligible (Çalışkan & Elçi, 2009). The climate of the study area is typical Mediterranean with moderately cold, rainy winters and hot, dry summers. The long-term mean annual precipitation is 690 mm.

Based on analysis of satellite imagery, the land use distribution of the area is 42.1% forests, 31.8% agricultural areas, 3.1% water bodies, 1.8% residential and 0.2% industrial areas According to the 2008 census data (Turkish Statistics Institution, 2008), approximately 68,000 inhabitants live within the boundaries of the stream basin. The population density is much higher closer to the city center (north of the basin). The basin area is subject to increases in population since the 1990's owing to agricultural activities and immigration from neighbouring districts. The study area is also under environmental stress, in particular with respect to groundwater. Some communities in the area rely on groundwater from supply wells. On the other hand, there are many wells drilled in the surficial aquifer, which provide the much needed irrigation water for agriculture. Excessive withdrawal from the surficial aquifer due to increase in population and also periodic droughts that have become more pronounced and sustained due to climate change pose a serious threat to the quantity and quality of groundwater resources.

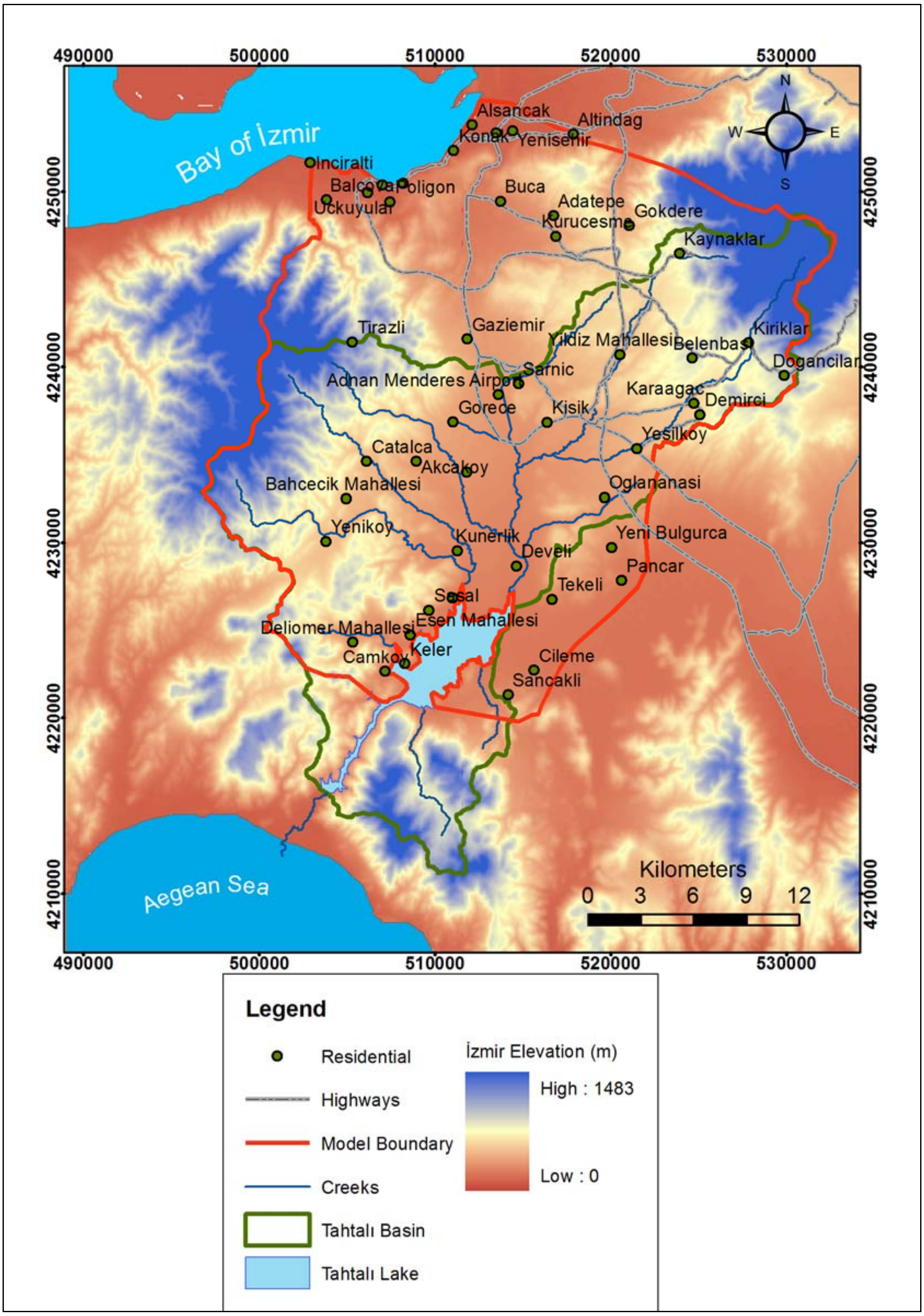

Figure 3.1 General location map of the study area

3.2 Hydrology of the Study Area

Various creeks and streams and their tributaries in the basin feed the Tahtalı dam reservoir. Namely these creeks are Tahtalıçay (Tahtalı stream), Şaşal, Deliömer, Balaban, Sandı, Sarıçay, Değirmendere, Kocaçay and Kona creeks. All of these streams are ephemeral in nature and go dry in the summer. The major water contribution to the reservoir is from the north via the Tahtalı and Şaşal streams. The annual average discharge of the Tahtalı stream is 152.3 million m^3. The discharges from the remaining inflowing streams are relatively negligible. Important springs in the study area are located in the northeastern part, near the Nif mountain. These are Ayrancılar (~0.9 m^3/s), Oğlananası (~0.25 m^3/s) and Gürlek (~0.5 m^3/s) (Şimşek, et.al, 2008).

The Tahtalı dam reservoir was constructed in the 1990s and is in service since 1997 to meet the water demand of the Izmir metropolitan area. It is a significant surface water resource which is designed to supply 3.5 m^3/s. However, the actual withdrawal rate from the reservoir varied due to operating issues and periodical droughts (Çalışkan & Elçi, 2009). The variation of the reservoir water level between the years 2000-2007 and available monthly precipitation data are shown in Figure 3.2. In general, the level of the reservoir decreases during the dry summer periods. The level reached its maximum in the winter season of 2003 season. It decreased to a minimum in 2007, which was an exceptional dry year with an annual total precipitation of only 487 mm.

Drinking water to the city of Izmir is supplied from various other resources, such as the deep wells Sarıkız, Göksu, Halkapınar, Çamdibi, Pınarbaşı, Menderes, Menemen and Çavuşköy, which supply about 133.6 km^3/year (IZSU, 2008). The Pınarbaşı and Menderes deep wells are located within the study area boundaries. The Pınarbaşı well site consists of two wells, which withdrew 679,132 and 894,827 m^3/year of groundwater for the years 2006 and 2007, respectively. The Menderes well site consists of nine alternately operating wells with a 20 L/s designed pumping rate for each well. All of these wells are exclusively operated by IZSU, the water authority of municipality of Izmir.

Numerous other smaller wells exist within the boundaries of the studied area. Most of these wells are used for irrigation of crops. In areas where the groundwater table is shallow these wells are in the form of dug wells. The wells in the area are partly registered at the regional directorate of the State Hydraulic Works (DSİ). Information and data regarding their depths, specific capacities, static and dynamic levels, and borehole logs are only available at registered wells. Other wells in the industrial area of Sarnıç are production wells for industrial process water. Some drinking water wells are located in the towns of Yeni Bulgurca, Develi, Kaynaklar, Görece, Kırıklar, Çileme, and Kısık, which were not operated by IZSU, but by the townships.

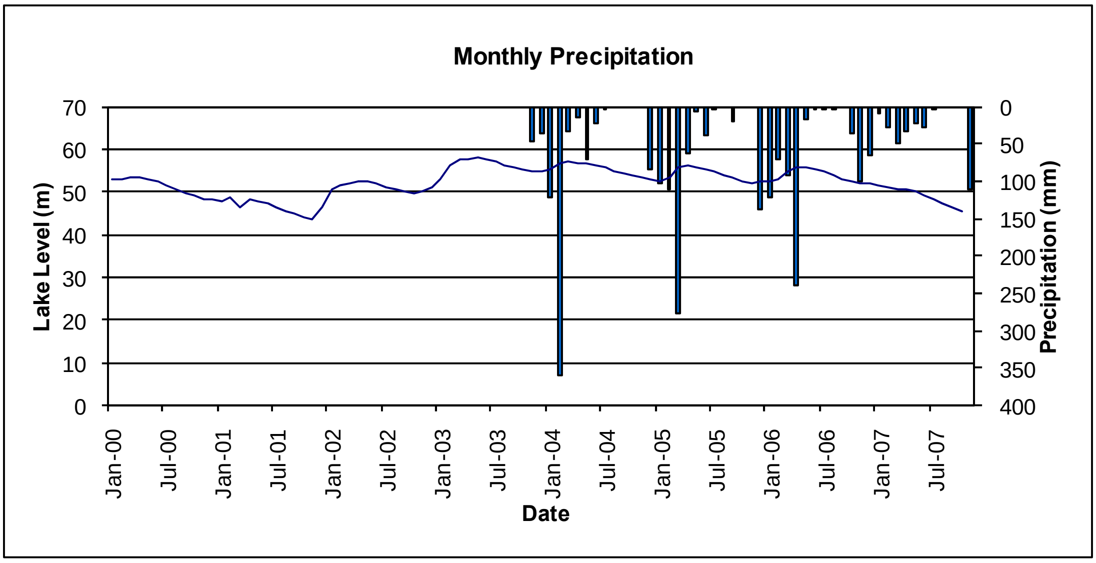

Figure 3.2 Monthly total water level records of the Tahtalı reservoir and monthly precipitation

3.3 Geological Setting

The geological properties of the study area were obtained from the 1/100,000 scale geological map, courtesy of the regional directorate of Mineral Research and Exploration (MTA), and also from previous regional studies and reports about the area (e.g. Koca, 1995). In general, quaternary aged alluvial deposits, neogene aged flysch, clayey limestone, allochthonous limestone, conglomerate and tuff formations comprise the geological structure of the study area (Figure 3.3). Alluvial deposits are dominant north of the Tahtalı reservoir extending up to the city of Izmir. The thickness of this formation can be as much as 80 m. At higher elevations this formation becomes thinner. The permeability of the alluvial is relatively high. Other unconsolidated sediments in the study area is a strip of marl-clay-tuff conglomerates in the west. However, most of the study area comprises of consolidated formations. These are limestone, flysch, marble, clayey schist and tuff. These formations can have very low permeabilities, with the exception of allochthonous limestone formation that occur in the northeast of the study area. These formations have a maximum thickness of about 200 m, and they are classified as the most important karstic formation in the region. These karstic rocks are surrounded primarily by flysch formations.

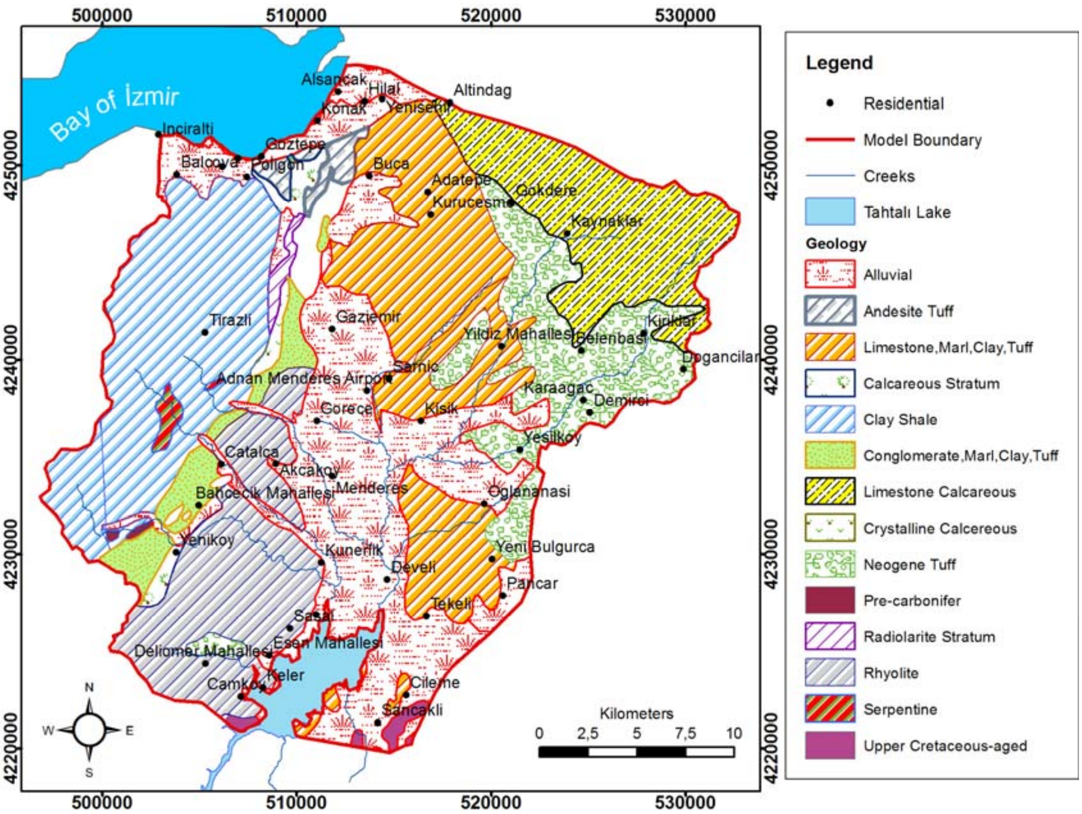

Figure 3.3 Geological map of the study area

CHAPTER FOUR
METHODS AND MODELS

4.1 Field Work

Any modeling study in particular groundwater flow modeling studies require extensive and accurate data to obtain reliable and useful results. Some data such as meteorological data (precipitation, evapotranspiration and temperature), flow rates for the Tahtalı stream and borehole logs to identify subsurface profiles were available. However, groundwater table elevations were not available and therefore field work was undertaken to measure the groundwater head in the surficial aquifer. Field work was performed three times during the duration of the study; the field trips occurred in February, May and October of 2007. Exploration of the study site was commenced in February 2007. The main goal was to find accessible wells that can be used to measure the water table elevation. A fairly homogeneous spatial distribution of these monitoring wells was also desired. The coordinates of every potential monitoring well was recorded with a GPS device (Figure 4.1). After the first field trip, different types of wells were encountered (Figures 4.2 and 4.3). All potential monitoring wells were tabulated and a short-list of the best wells was established. All relevant properties of these selected 51 monitoring wells were then put into a database (Table 4.1). Later, in May and October of 2007, field trips were organized to measure the groundwater elevations. The measurements conducted in May represented the wet season and the measurements in October the dry season, which occur immediately after the hot and dry irrigation season. During the field trip in October, it was observed that some wells went dry or became inaccessible for measurement. Therefore, these wells were substituted by other wells in the vicinity to serve as monitoring points.

The depth to the water table in each monitoring well was measured using a 150m-length electrical tape water level meter manufactured by Akım Elektronik Ltd.Şti.. It has a stainless steel probe that contains two electrodes with an insulating gap between them. A buzzer sounds when the probe makes contact with water. The depth to the water table is then read from the measuring tape that is connected to the probe.

Groundwater elevations were computed by substracting the measured groundwater depth from the ground elevation (Z). Calculated groundwater elevations are given in Table 4.1.

Figure 4.1 Fixing the coordinates of a dug well with a GPS during the field trip in May 2007

Besides groundwater elevation measurements, the geological properties and, if possible, soil profiles (Figure 4.4) in the study area were observed and recorded to verify the accuracy of the geological map. The original geological map was revised accordingly to match field observations.

Figure 4.2 Different types of wells were selected for groundwater level measurements: deep irrigation well with an electrical submergible pump (above) and a shallow well with a lift pump (below)

Figure 4.3 Dug wells used as monitoring wells

Figure 4.4 Soil profiles in the study area: alluvial in the left photo and conglomerates in the distance of the right photo

Figure 4.5 Measurement of the groundwater depth with a water level meter

Table 4.1 Monitoring well database

Well no	X (m)	Y (m)	Elevation, Z (m)	Groundwater depth (m)-May	Groundwater level (m)-May	Groundwater depth (m)-October	Groundwater level (m)- October	Well Description
K12	516931	4247575	145	52.96	**92.04**	53.17	**91.83**	Deep supplying well
K14	520575	4248364	168	3.99	**164.01**	5.81	**162.19**	Dug well
K18	512247	4236651	137	7.60	**129.40**	8.76	**128.24**	Deep irrigation well
K25	523465	4236389	201	55.05	**145.95**	50.64	**150.36**	Dug well
NKT007	516097	4225548	58	18.20	**39.80**	24.98	**33.02**	Deep irrigation well
NKT010	514170	4222203	71	2.90	**68.10**	5.73	**65.27**	Dug well
NKT015	514467	4224023	55	9.87	**45.13**	5.73	**49.27**	Deep irrigation well with centrifugal pump
NKT019	513962	4228481	71	5.97	**65.03**	7.74	**63.26**	Dug well
NKT021	512716	4227294	78	10.60	**67.40**	18.43	**59.57**	Artesian well
NKT022	512199	4226823	72	6.45	**65.55**	13.04	**58.96**	Deep supplying well
NKT025	509807	4226045	93	2.05	**90.95**	3.03	**89.97**	Şaşal deep drinking well
NKT027	510902	4229536	92	11.15	**80.85**	13.74	**78.26**	Artesian Well
NKT029	514992	4229239	76	5.70	**70.30**	unmeasured	**n/a**	Artesian Well
NKT030	514089	4230198	84	4.94	**79.06**	32.78	**51.22**	Deep drilling well with submergible pump
NKT034	512792	4231945	91	6.79	**84.21**	20.06	**70.94**	Deep irrigation well
NKT036	511326	4231667	99	11.19	**87.81**	15.75	**83.25**	Deep irrigation well
NKT037	511382	4232620	105	6.44	**98.56**	6.88	**98.12**	Deep irrigation well
NKT039	511015	4233566	122	2.33	**119.67**	unmeasured	**n/a**	Dug well for irrigation
NKT041	512230	4233233	108	2.67	**105.33**	1.64	**106.36**	Deep supplying well
NKT042	513289	4232957	100	7.19	**92.81**	unmeasured	**n/a**	Dug well
NKT047	525558	4240763	323	8.47	**314.53**	9.28	**313.72**	Dug well
NKT048	519916	4245788	244	73.06	**170.94**	73.75	**170.25**	Artesian well
NKT051	523234	4244791	265	9.37	**255.63**	26.30	**238.70**	Deep irrigation well
NKT052	522828	4243616	253	3.43	**249.57**	4.43	**248.57**	Deep supplying well
NKT055	526245	4239243	267	4.64	**262.36**	4.78	**262.22**	Deep irrigation well
NKT057	521407	4235258	171	12.42	**158.58**	8.44	**162.56**	Deep supplying well
NKT058	518656	4233149	130	9.00	**121.00**	unmeasured	**n/a**	Municipal well

Table 4.1 (continued) Monitoring well database

NKT060	521535	4227470	54	2.28	**51.72**	75.76	**-21.76**	Deep irrigation well
NKT062	514755	4234711	100	3.28	**96.72**	6.21	**93.79**	Dug well
NKT063	510522	4235833	142	10.80	**131.20**	11.33	**130.67**	Dug well
NKT064	507675	4233002	147	2.79	**144.21**	5.87	**141.13**	Artesian well
NKT065	505242	4231435	176	2.04	**173.96**	2.80	**173.20**	Artesian well
NKT066	512065	4241000	135	9.97	**125.03**	10.11	**124.89**	Deep supplying well
NKT067	524514	4240998	362	9.00	**353.00**	22.51	**339.49**	Deep supplying well
NKT068	524631	4237838	229	21.86	**207.14**	22.35	**206.65**	Dug well for irrigation
NKT070	517731	4235913	130	9.18	**120.82**	18.59	**111.41**	Deep supplying well
NKT071	514386	4232121	91	3.42	**87.58**	7.62	**83.38**	Dug well
NKT072	513757	4233929	105	8.34	**96.66**	11.56	**93.44**	Dug well
NKT073	510753	4233153	109	3.68	**105.32**	7.19	**101.81**	Dug well
NKT074	521179	4238200	125	26.61	**98.39**	42.84	**82.16**	Deep irrigation well
NKT075	512391.1	4227942	83	18.79	**64.21**	unmeasured	**n/a**	Dug well
NKT076	514985.2	4229060	71	8.51	**62.49**	unmeasured	**n/a**	Drilling well
NKT078	512950.2	4221388	84	5.12	**78.88**	5.79	**78.21**	Artesian well
NKT079	511832.1	4220924	79	3.62	**75.38**	unmeasured	**n/a**	Deep supplying well
NKT080	517996.1	4224184	52	24.60	**27.40**	28.00	**24.00**	Drilling well
NKT082	514702.4	4249034	107	28.10	**78.90**	28.69	**78.31**	Artesian well
NKT083	523636.2	4246211	283	6.61	**276.39**	8.10	**274.90**	Dug well
NKT084	528391.6	4240473	336	4.90	**331.10**	unmeasured	**n/a**	Drilling well
NKT085	522992.4	4237580	216	26.55	**189.45**	34.09	**181.91**	Artesian well
NKT086	521755.7	4237257	199	36.28	**162.72**	unmeasured	**n/a**	Artesian well
NKT087	520763.4	4241086	183	4.56	**178.44**	5.71	**177.29**	Drilling well
NKT090	511958.9	4221099	84	unmeasured	**n/a**	22.86	**61.14**	Dug well
NKT091	513516	4222814	70	unmeasured	**n/a**	6.41	**63.59**	Dug well
NKT092	527348.2	4240053	303	unmeasured	**n/a**	5.92	**297.08**	Dug well

4.2 Modeling Approach

4.2.1 Groundwater Flow Model Setup and Execution

The groundwater flow equations that comprise the groundwater flow model of this study were solved using the model code MODFLOW-2000 that is based on the finite-difference method. The groundwater flow model was set up as a one-layered, regional steady-state model. The purpose of the model was to simulate groundwater flow of the unconfined aquifer, and thereby calculate the distribution of water table elevations and groundwater fluxes. The extent of the modeling domain was the same as the extent of the study area boundaries shown in Figure 3.1. The boundaries of the model were determined such that it encompasses the entire area of interest and coincides with hydrological boundaries, e.g. sea, lake, watershed boundaries. The modeling domain was divided up (discretized) into equal-sized 150×150 m finite-difference grid cells. Furthermore, more than 100 borehole logs were processed to determine the depth to the impermeable layers, which were subsequently interpolated to obtain the surface representing the bottom surface of the model layer. The top surface of the model was obtained directly from 90m-resolution SRTM data (NASA'a Shuttle Radar Topography Mission). Details about the boundary conditions and the spatial discretization of the model are discussed in the next chapter. Other secondary model input parameters were the extraction rates of major agricultural, domestic and industrial water supply wells in the study area and the water surface elevations of streams.

4.2.2 Model Calibration and Verification

Hydraulic conductivity and vertical groundwater recharge from precipitation were the key parameters of the model. In this study, recharge was considered as net recharge, i.e. the actual portion of water reaching the water table after being withdrawn by plants in the root zone, thereby eliminating the need for the evapotranspiration parameter. These model parameters were handled as calibration parameters, which were varied within a plausible range of values during the calibration process. Calibration of the model was performed automatically using the parameter estimation code PEST (Doherty, 2004). The aim of the calibration process was to adjust the calibration parameters in a systematic manner in order to obtain a satisfactory match between measured water table elevations and the calculated values by the model. The model was pre-calibrated manually on a trial-and-error basis before the automatic calibration procedure with PEST was initiated. Thereby, an optimum starting point was achieved for the automatic calibration, which resulted in a more robust performance of the parameter estimation process with PEST.

The model domain was split up into six hydraulic conductivity zones, each zone representing a different geological formation and thus expected to have different but uniform hydraulic conductivities. Each zone was handled as a separate calibration parameter, which were varied within a plausible range of values during the calibration process. The plausible range of hydraulic conductivities was known apriori based on the properties of the different formations published in the literature (Fetter, 2001; Spitz & Moreno, 1996). Similarly, the model domain was divided into recharge zones, in each of which recharge originating from the ground surface was assumed to be uniform. The recharge rate for each zone was unknown and thus was used as calibration parameters of the model, except for one zone, which represents the Tahtalı watershed. For this particular recharge zone, recharge was calculated with an independent, transient regional-scale precipitation-runoff model (Fıstıkoğlu & Harmancıoğlu, 2001). The details of this model are explained in the next section.

The model was calibrated to the wet season water table depths measured in May 2007 at 51 wells, and subsequently the calibrated model was verified with an independent data set representing the dry season water table measured in October 2007 at the same wells. During the verification of the model, the boundary conditions and recharge rates were modified accordingly to match wet season conditions. For example the water level of the Tahtalı reservoir, which was considered as a constant-head boundary condition in the model, was adjusted to the summer seasonal average. The average reservoir level was 51.18 m for the winter period of November 2006-April 2007. This value was used during the calibration of the model. It was then reduced to 47.53 m, which represented the average level for the summer period of

May-October 2007, for the verification run of the model. Hydraulic conductivity values remained unchanged and the same values from the calibrated model were used in the verification of the model.

To evaluate the performance of the model, some fundamental calibration statistics were determined. In this study, the residual was defined as the measured water table elevation minus the calculated. The residual mean, absolute residual mean, residual standard deviation (or root mean squared deviation, RMSD), sum of residual squares and the ratio of RMSD to the range in measured values were determined. The distribution of residuals were also plotted to assess the possible presence of systematic errors in the model.

4.2.3 Precipitation-Runoff Modeling to Estimate Groundwater Recharge

Groundwater recharge is one of the key parameters of a groundwater flow model. Different methods exist to estimate groundwater recharge. These methods can be divided into physical, chemical (tracer) and numerical modeling approaches. numerical models are useful and robust tools to quantify recharge. Precipitation-runoff modeling or sometimes referred to as watershed modeling is a surface-water focused approach, which generally yield groundwater recharge estimates as a residual term in the water budget equation (Wanke, Dünkeloh & Udluft, 2008). These models are usually lumped and provide a single recharge estimate for the entire watershed.

In this study, the entire Tahtalı stream basin was treated as a single recharge zone. The remaining areas within the study area were divided into separate recharge zones. The recharge rate for the Tahtalı basin was determined using a water budget model that estimates the monthly surface water flow rates at the lowest pour point of the basin using monthly precipitation and evapotranspiration records. It also includes the calculation of the components, actual evapotranspiration, interflow, baseflow, subsurface storage and percolation, the latter was used in this study as an input parameter of the groundwater flow model. As opposed to distributed models, all parameters of the water budget model were considered to be averaged and representative of the entire basin, which was later defined as a separate recharge zone in the groundwater flow model. The flow chart and equations used in the model are shown in Figure 4.6. The parameters and variables used are as follows:

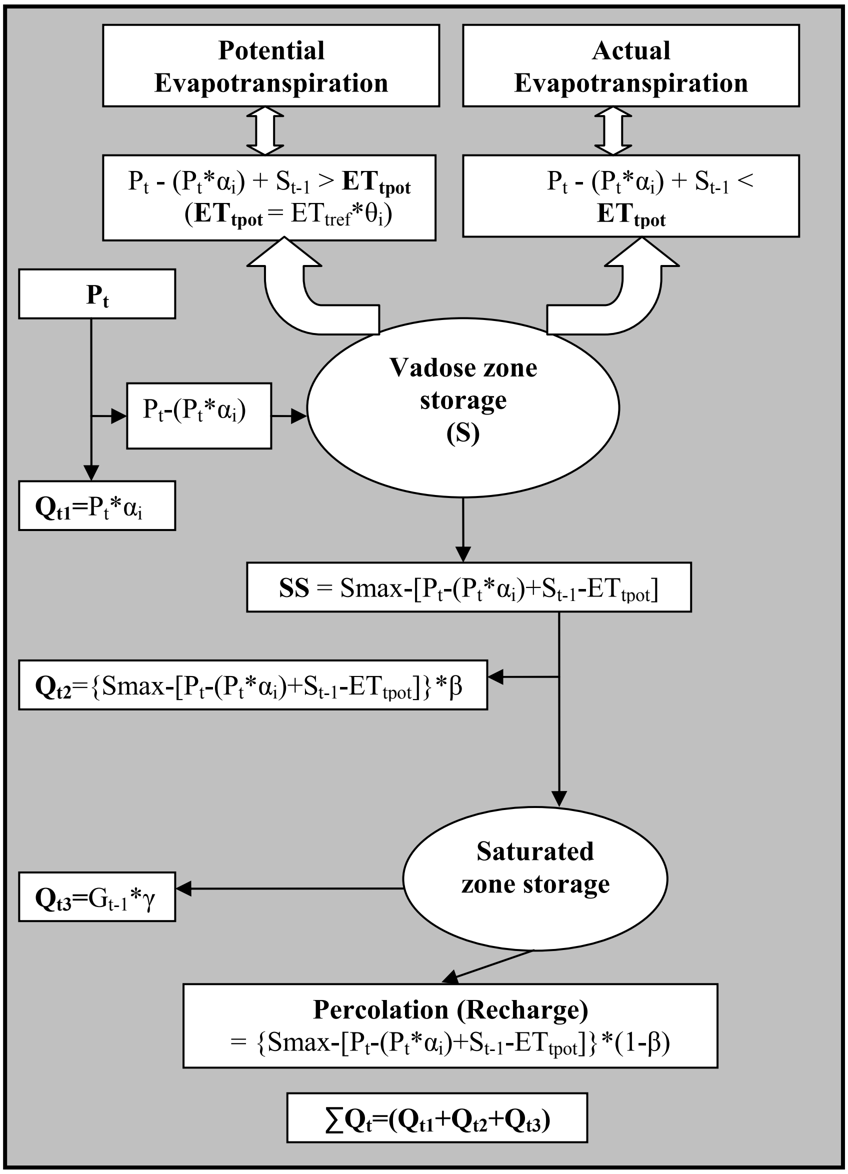

Figure 4.6 Components and flow chart of the water budget based precipitation-runoff model (adapted from Fıstıkoğlu & Harmancıoğlu (2001)

P_t: total precipitation for month t (mm/month)

ET_{tref}: monthly total reference evapotranspiration (mm/month)

ET_{tpot}: monthly total potential evapotranspiration (mm/month)

θ_i: coefficient dependent on vegetation type

α_i: runoff coefficient dependent on land use and vegetation type

β: interflow coefficient

γ: baseflow coefficient

Smax: Maximum storage capacity of the soil (mm)

SS: vadose zone storage surplus (mm)

S_t: vadose zone storage for month t (mm)

G_t : saturated zone storage for month t (mm)

Q_{t1}, Q_{t2}, Q_{t3}: surface runoff, interflow, baseflow

ΣQ_t: total surface flow rate at the basin pour point for month t (mm)

P_t and ET_{tref} are the required input parameters that were obtained from meteorological monitoring stations. Using the relationships between the several compartments, ΣQ_t was successively calculated by the model for each month. The remaining coefficients were the model's calibration parameters and were varied within defined ranges until a satisfying match between ΣQ_t and the observed surface flow rates of the Tahtalı stream was obtained. Observed flow rates were available for October 1969-September 1988. The model was calibrated to the observed flow rates of October 1969 to September 1981. Linear regression and Nash-Sutcliffe coefficients were used as criteria to evaluate the performance of the model. The calibrate model was then verified using the data from the period October 1981 to September 1988. Finally, the calibrated model was run to calculate stream flow rates for the time period of October 1988 to May 2008.

Recharge calculated by this model was averaged in time to yield appropriate input for the steady-state groundwater flow model. Calculated recharge values for the period November 2006-April 2007 was averaged and used as input for the recharge zone that represents the Tahtalı watershed in the groundwater flow model. Similarly, the output for May 2007-October 2007 was representative of the dry period and hence was used as input for the verification of the calibrated groundwater flow model.

4.3 Other Software Used

Several commercially available software packages were used during the development of the groundwater flow model in this study. Some of these packages such as RockWorks and Surfer were mainly used as processing tools to generate model input data from available information and data obtained from field work. The main task of this study, which is the numerical modeling of groundwater flow for the study area, was carried out utilizing the groundwater modeling package software Groundwater Vistas v4.0. Other software like ArcGIS v.9.1 and Microsoft Excel were necessary for the post-processing of modeling results, and transform them into visually presentable outputs. Brief software descriptions and the extent they were utilized are provided in this section.

4.3.1 Groundwater Flow Model MODFLOW-2000 within Groundwater Vistas

Groundwater Vistas (GWV) is a Microsoft Windows based, sophisticated graphical user interface for comprehensive 3-D groundwater flow and contaminant transport modeling tasks. It couples a model design system with elaborated graphical analysis tools. GWV integrates many public-domain and commercial hydrogeological models, such as MODFLOW, MODPATH (both steady-state and transient versions), MT3DMS, MODFLOWT, MODFLOW-SURFACT, GFLOW, RT3D, PATH3D, SEAWAT and PEST, the automatic calibration code. (Spink, 2007)

4.3.2 RockWorks 2006

RockWorks 2006 developed by Rockware, Inc. is an integrated software package for geological data management, analysis, and visualization. It specializes in visualization of subsurface data in the form of logs, cross-sections, fence diagrams, solid models, structural and isopach maps both in 2-D and 3-D. Rockworks 2006 manual, 2008) RockWorks analyzes subsurface data such as stratigraphy, lithology, fracture data, hydrology and geochemical data. In this study, it was used to determine the depth of the impermeable layer under the surficial aquifer and thereby the thickness of the model layer. More than 100 borehole logs in hardcopy format that contained information on geographical location, lithology, stratigraphy, pumping tests, and in some cases static and dynamic water table depths were put into the RockWorks database and processed to obtain the depth to the impermeable layer at each given borehole location. The information was further processed with Rockworks to obtain a continuous surface representing the bottom of the surficial aquifer that is assigned as the bottom surface of the model layer.

4.3.3 ArcGIS

A geographical information system (GIS) can be defined as a system for entering, storing, manipulating, analyzing and displaying geographic or spatial data. ArcGIS v.9.1 developed by ESRI, Inc. was used in this study to store all relevant data and produce miscellaneous maps. Information about the study area, such as locations of towns/villages, creeks/streams, highways etc. were digitized from a scanned 1/25000 topographical map. Similarly, the geological map of the study area was digitized from a hardcopy version of the map with the help of this software.

CHAPTER FIVE
MODEL DESCRIPTION

5.1 Spatial Discretization

The model domain was divided into a number of discrete finite-difference grids. A 150×150 m, cell-centered finite-difference mesh with 33728 grid cells was used. In the vertical dimension, the model was single-layered and the top elevation surface of the model represented the land surface of the study area. Using 90m-resolution SRTM data top elevation values for each grid cell was assigned. The unconfined aquifer with the study area boundaries was modeled as one layer, with the MODFLOW setting 'LAYCON=1, unconfined'. The layer's bottom elevation represented the confining layer of the unconfined aquifer. The bottom elevations were determined by evaluating the stratigraphic information in well logs; the depth values at each well log location to the impermeable layer below the unconfined aquifer were interpolated on a surface, which was assigned as the bottom elevation surface of the model layer. The resulting thickness of the model layer is shown in Figure 5.1.

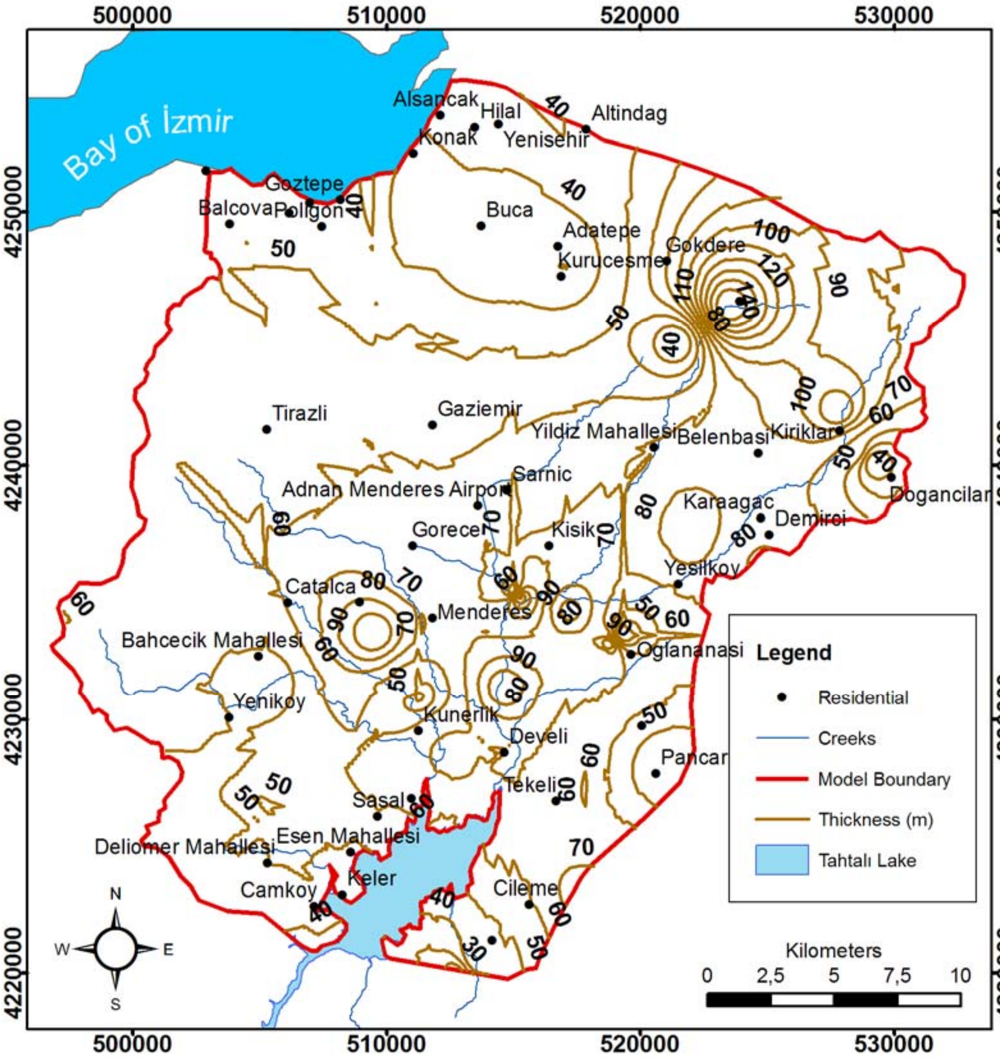

Figure 5.1 Isopach map showing the thickness of the model layer

5.2 Model Parameters

5.2.1 Hydraulic Conductivity

One of the key model parameters was the hydraulic conductivity. It is a measurable property of any aquifer system, however, sufficient measurements were not available for the study area. Therefore, it was handled as an uncertain parameter that was determined using the calibration process. Since the groundwater flow model was single-layered, only horizontal (lateral) hydraulic conductivities were relevant. The model domain was divided into six hydraulic conductivity zones (Figure 5.2). Each zone represented a different a different geological formation or a group of formations with similar properties. Therefore, each zone was assigned different but uniform hydraulic conductivity values. Initial hydraulic conductivity values were obtained from literature according to aquifer properties. These values were varied within a pre-defined plausible range during the calibration process. Initial hydraulic conductivity values and the calibration ranges are presented in Table 5.1.

Table 5.1 Initially assigned hydraulic conductivity values and calibration ranges

Hydraulic conductivity zone	**Initial Value (m/d)**	**Lower Bound (m/d)**	**Upper Bound (m/d)**
K_{xy}-1 (Alluvial)	30	1	100
K_{xy}-2 (Allochthonous Limestone	0.03	0.01	10
K_{xy}-3 (Flysh)	0.3	0.001	1
K_{xy}-4 (Tuff)	0.3	0.1	10
K_{xy}-5 (Conglomerate)	1	0.05	5
K_{xy}-6 (Clayey Limestone)	1	0.05	5

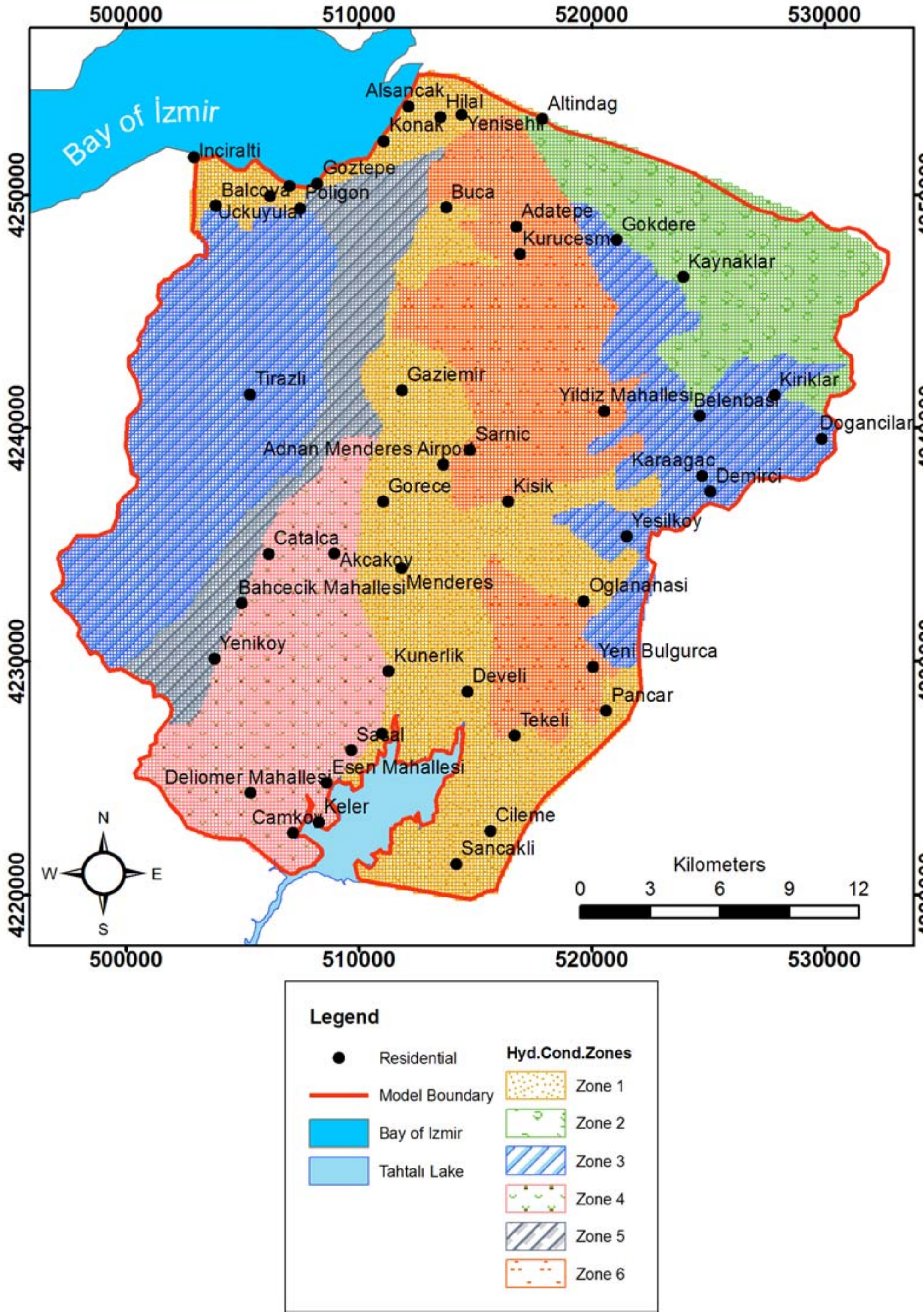

Figure 5.2 Hydraulic conductivity zones used in the model

5.2.2 Recharge

In this study, recharge was considered as net recharge, i.e. the actual portion of water reaching the water table after being withdrawn by plants in the root zone, thereby eliminating the need for the evapotranspiration parameter. Similar to the approach in hydraulic conductivity, the model domain was divided into four zones, representing recharge areas that are likely to have recharge rates in the same order of magnitude. In Figure 5.3 recharge zones used in the model are shown. Recharge zone number 4 represented the entire Tahtalı stream watershed. For this particular recharge zone, recharge was calculated with an independent, transient regional-scale precipitation-runoff model (Section 4.2.3). The recharge rate calculated by this model was assigned to recharge zone 4. Recharge rates for the other zone were largely unknown and therefore handled as calibration parameters. Recharge zone 1 covered the urban part of the city area with mostly an impervious ground surface that would generate relatively more runoff. Recharge zone 2 represented the foothills of the Nif mountain, which consist mostly of limestone. The alluvial located outside the Tahtalı stream basin east of the Tahtalı reservoir was assigned as recharge zone 3. Initial values for these zones were estimated based on precipitation records for the period November 2006 – April 2007. The total amount of precipitation during this period was 234.6 mm, which translates to an average daily precipitation rate of 1.30×10^{-3} m/d. It was assumed that 1/3 of the precipitation ($4.34\ \times10^{-5}$ mm/d) recharged the aquifer percolated to the water table. The initial recharge rate for each recharge zone was assigned values in this order of magnitude. The lowest and highest initial recharge zone rate was assigned to Recharge zone 1 and 2, respectively. The recharge rate assignments and the calibration ranges are shown in Table 5.2.

Table 5.2 Initial recharge rates and calibration ranges

Recharge zone	Initial Value (m/d)	Lower Bound (m/d)	Upper Bound (m/d)
Zone 1 - R1	3×10^{-5}	5×10^{-4}	1×10^{-6}
Zone 2 - R2	7×10^{-5}	5×10^{-4}	1×10^{-6}
Zone 3 - R3	1.5×10^{-4}	5×10^{-4}	1×10^{-6}
Zone 4 - R4	9.016×10^{-5}	not calibrated	

The average daily recharge rate for zone 4 was calculated by the precipitation-runoff model as 9.016×10^{-5} m/d, which corresponded to a total recharge of 16 mm for the period of November 2006 - April 2007. Detailed results of the precipitation-runoff model are presented in the next chapter.

Recharge rates were modified for the verification of the groundwater flow model to accommodate dry summer conditions. It was assumed that the decrease in the recharge rate for the dry period is proportional to the decrease in precipitation. Based on meteorological records, the decrease in precipitation for the period of May 2007 – October 2007 was 12.5%. Therefore, the calibrated recharge rate for each zone in the model was decreased by that percentage. Resulting recharge rates for the verification period of the model are presented in Table 5.3.

Table 5.3 Modified recharge rates for the verification of the model

Recharge zone	Calibrated Value (m/d)	Modified recharge used for verification (m/d)
Zone 1- R1	6.27×10^{-5}	5.49×10^{-5}
Zone 2- R2	1.28×10^{-4}	1.12×10^{-4}
Zone 3- R3	5×10^{-4}	4.375×10^{-4}
Zone 4- R4	9.016×10^{-5}	7.889×10^{-5}

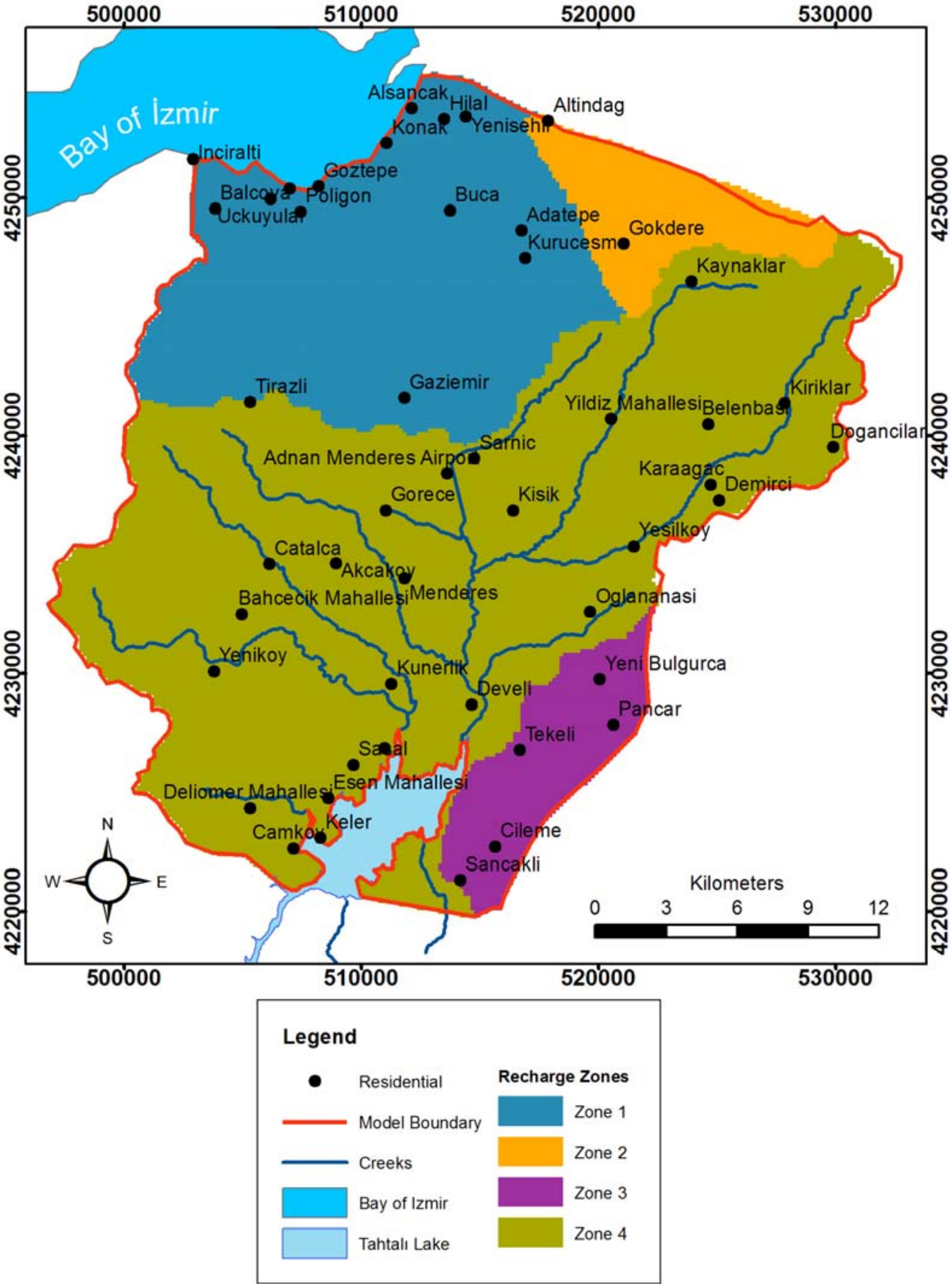

Figure 5.3 Recharge zones used in the model

5.2.3 *Pumping Wells*

Water production wells within the study area that are used for drinking, industrial and irrigation water supply were determined and major wells that withdraw groundwater more than 3 L/s were shortlisted to use as groundwater sinks in the model. A total of ten individual water supply wells and wellfields (Table 5.4) were considered in the model. A great deal of wells is called municipal drinking water wells. The discharge rates for wells with no available data were estimated based on the current population that they supply water, assuming a water consumption of 300 L/d/capita. Eventually, a total discharge rate of 25859.4 m^3 /d was assigned to all production wells in the model. The locations of these wells are shown in Figure 5.4.

Table 5.4. Water supply wells and well fields considered in the model

Well Name	**Pumping rate (m^3/d)**
Pancar drinking water well	1296
Oğlananası drinking water well	700.2
Kısık drinking water well	166.2
Develi drinking water well	575.7
Menderes drinking water wells	5037.6
Çileme drinking water well	375
Yeni Bulgurca drinking water well	675
Sarnıç industrial water supply wells	13937.1
Şaşal drinking water well	555
Kırıklar drinking water well	312.3

5.2.4 Surface Water Interaction by Streams and Creeks

Streams and creeks in the study area such as Değirmendere, Kocaçay, Kona, Tahtalı, Sarıçay, Sandı, Balaban and Deliömer were considered in the groundwater flow model using the ‘river package’ of MODFLOW-2000. This package is a subroutine of the modeling code which was developed on the principle that the water exchange between the stream and the groundwater is a function of the head differential, the thickness, width and hydraulic conductivity of the streambed. This implies that the model gridcells that are assigned as streams can either act as groundwater source or sink, mainly depending on the head difference between the water table elevation and the water level of the stream. The reader is referred to McDonald & Harbaugh (1988) for the fundamentals behind this particular MODFLOW package and further details.

All streams were digitized from the base map of the study area using the GIS software. Only portions of the streams that were observed to have noticeable flow were included in the model. These are presented in Figure 5.5. Since the streams are ephemeral and the water depths are typically less than a meter, the stream water levels were assumed to be equal to the ground elevation. Thicknesses and hydraulic conductivities were lumped into one model parameter, which is referred to as streambed conductance. Streambed conductances in the order of magnitude of 10^3 m^2/d were manually adjusted during the manual pre-calibration of the model.

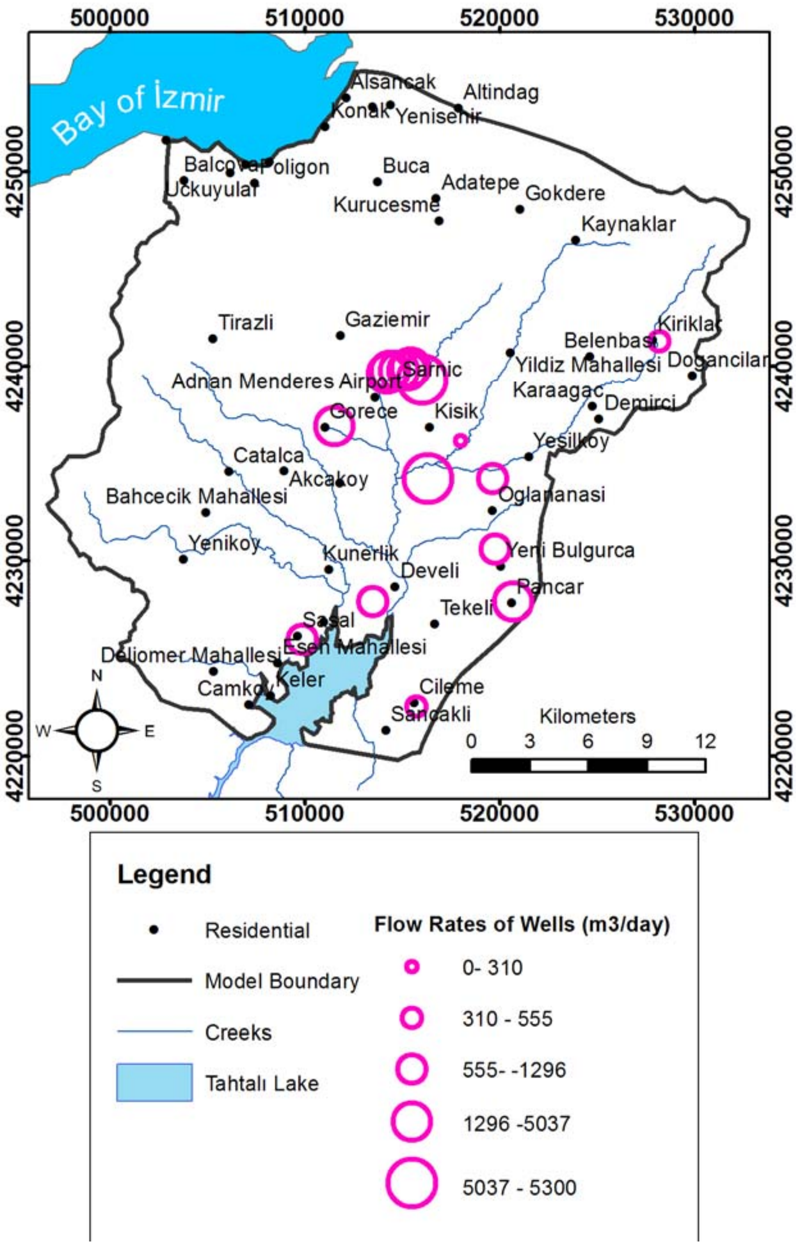

Figure 5.4 Water supply wells included in the model

5.3 Boundary Conditions

Boundaries of the groundwater flow model were defined as geological and hydrological features that influence the pattern of groundwater flow such as watershed boundaries, faults, surface water features, outcrops and water table divides. Model boundaries and the types of boundary conditions are shown in Figure 5.5. The boundaries of the model coincided mostly with the boundaries of the Tahtalı stream basin with the exception in the north and the southeast of the model domain. All three types of boundary conditions, namely constant-head (Dirichlet), defined-flux (no-flow) and mixed-type (Cauchy), were applied in this study. Boundary conditions used in the model are discussed in the following sections.

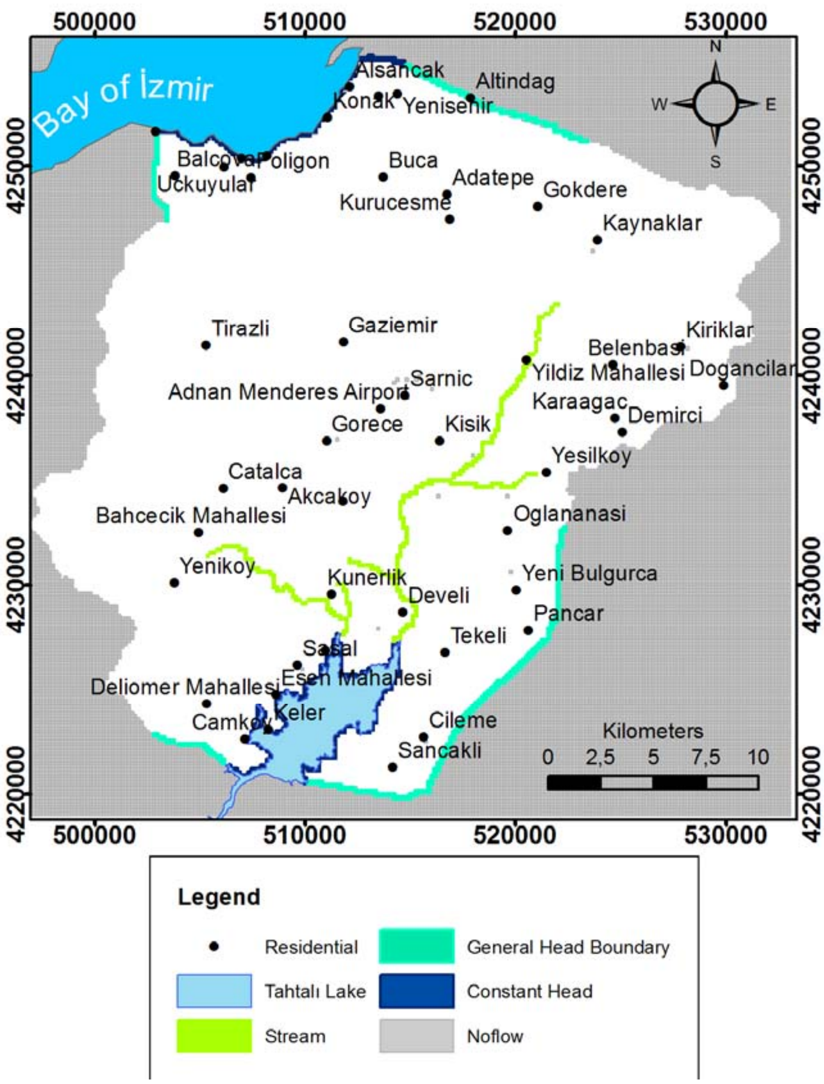

Figure 5.5 Boundary conditions used in the model

5.3.1 No-Flow Boundaries

Model boundaries that coincide with watershed boundaries were defined as no-flow boundaries. For example, the no-flow boundaries in the west of the model domain coincide with western boundaries of the Tahtalı and Balçova watershed boundaries. Similarly, the no-flow boundary in the east coincides with the Tahtalı watershed boundary. The boundary in the north of the model domain is based on the topographical divide of the Nif Mountain, with the assumption that the groundwater divide coincides with this topographical divide.

5.3.2 Constant-Head Boundaries

The Bay of Izmir in the north and the Tahtalı reservoir in the south were considered as constant head boundaries. The Bay of Izmir's constant head value was assigned as zero, equal to the sea level. The constant head value assignment for the Tahtalı reservoir was based on the actual reservoir level measurements obtained from IZSU and differed depending on the season; in the calibrated model the boundary was set at 51.18 m, and in the verification run of the model the value was modified to 47.53 m.

5.3.3 General-Head Boundaries

General-head boundaries (GHB) are third-type boundary conditions, which are basically head-dependent. The water flux at the boundary is essentially a function of the lateral head gradient and a given conductance, similar to the conceptualization of the river package in MODFLOW, only that the GHBs affect groundwater flux in the lateral direction. Boundaries of the model that could neither are assigned as no-flow or constant head were defined as GHBs, which permitted groundwater flux interactions with neighboring aquifers.

GHBs require the assignment of two parameters; a conductance value and head value representing a distant constant-head boundary. Conductance values in the order of magnitude of 10^4 m^2/d were used and varied during the manual pre-calibration process. The GHB head values were initially estimated using head values that were measured in neighboring monitoring wells. They were also varied during the pre-calibration process. However, for the verification run of the model the GHB head values were reduced to accommodate dry summer conditions. The seasonal reduction rates of the water table depth measured at the nearest monitoring well were determined and were applied to the GHB head values. Conductances were unchanged in the verification run.

CHAPTER SIX
RESULTS

6.1 Estimation of Recharge for the Tahtalı Basin

The recharge rate for the recharge zone representing the Tahtalı stream basin was calculated by the precipitation-runoff model. The recharge rates for the remaining zones were handled as calibration parameters. This groundwater flow model was unconventional in the sense that the groundwater recharge parameter in the model was estimated using a lumped, transient water-budget based precipitation-runoff model that was executed independent of the groundwater flow model. The recharge rate obtained from the calibrated precipitation-runoff model was used as input to the groundwater flow model, which was eventually calibrated to measured water table elevations.

The calibration of the precipitation-runoff model yielded regression and Nash-Sutcliffe coefficients of 0.85 and 0.83, respectively. The match between modeled and observed stream flow rates was satisfactory. Calculated and observed stream flowrates for the calibration and verification periods are shown in Figures 6.1 and 6.2, respectively. When visually compared, the match is good except for the 1984-1985 winter period. Here the model overestimates the stream flow. Illustrated in Figures 6.1, 6.2 and 6.3 are also the observed precipitation rates compared to the modeled percolation, i.e. groundwater recharge rates. Based on the model results, it can be stated that recharge for the summer months was generally calculated as very low, whereas for the winter months the percentage of precipitation that reached the water table in the form of recharge in the rainy winter months of the period 2003-2008 varied between 0 to 55.3%, depending on meteorological conditions of each year.

The modeling results for the period from October 1988 to April 2008 are shown in Figure 6.3. Here it can be noted that modeled stream flow rates and percolation rates are significantly lower for the winter months of 2006-2007. During this period an extreme drought occurred, which was coincidentally also the period in which this study was conducted and the groundwater flow model was calibrated. Furthermore, the recharge ratio was calculated to be equal to zero for the unusually dry summer months that occurred in that same period. The recharge rate for the wet period of 2007 was 16 mm, which corresponded to an average recharge to precipitation ratio of 3.3%. This recharge rate was directly used in the groundwater flow model as a recharge parameter for recharge zone 4.

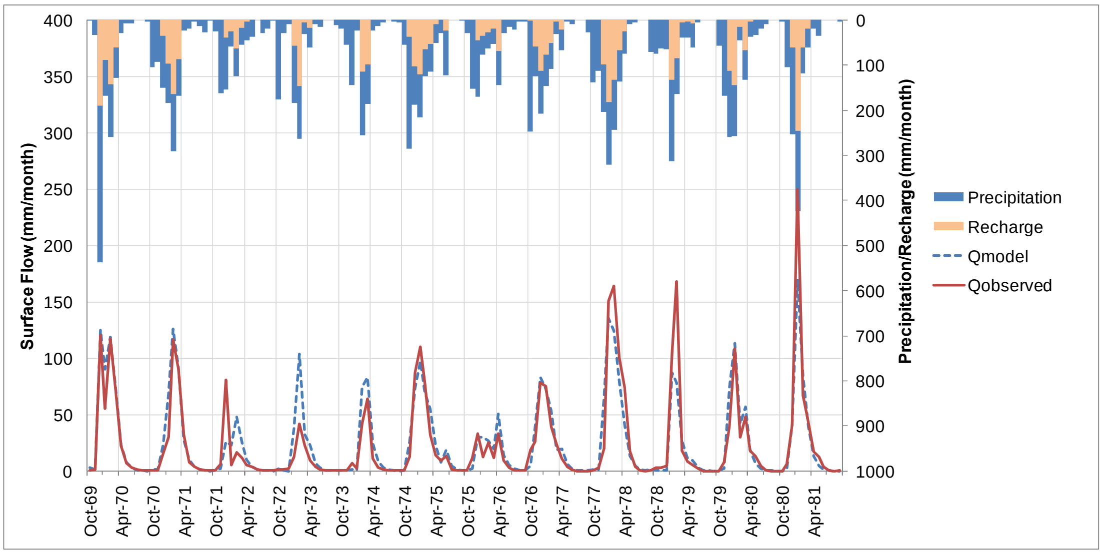

Figure 6.1 Results of the precipitation-runoff model; hydrographs for the verification period

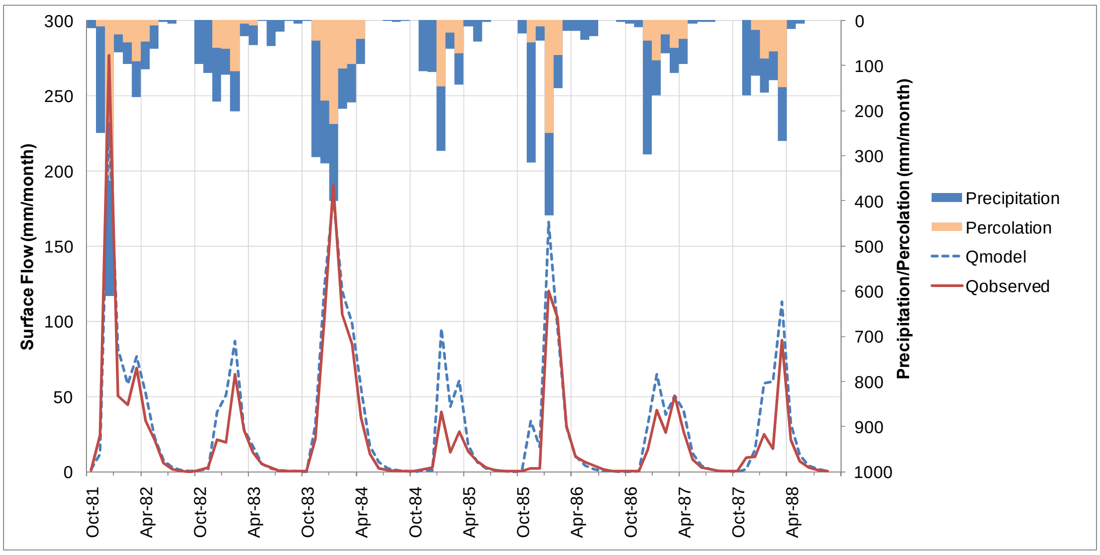

Figure 6.2 Results of the precipitation-runoff model; hydrographs for the verification period

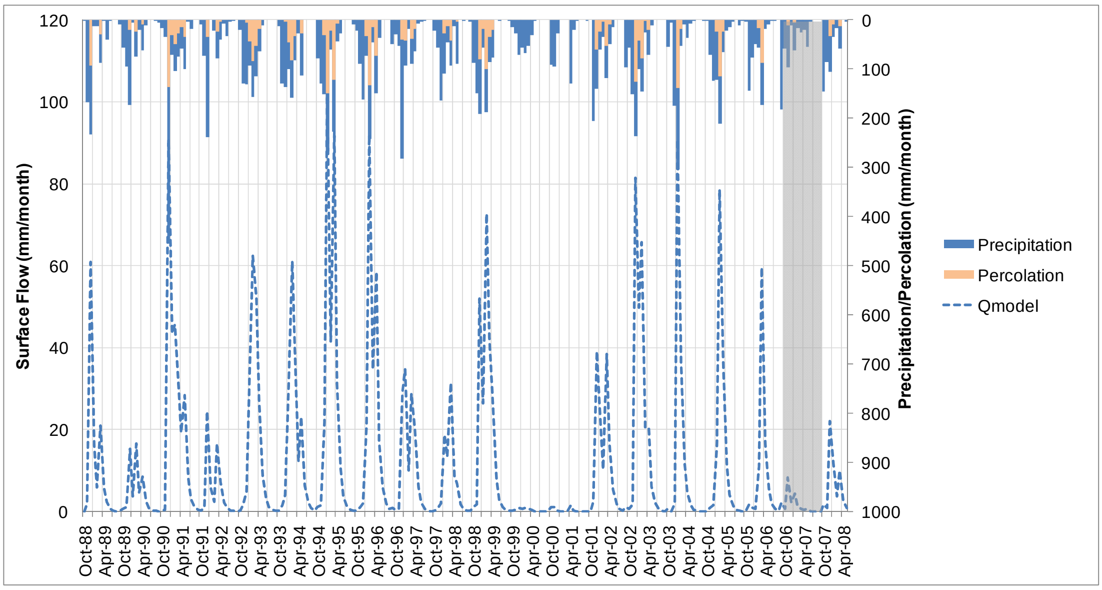

Figure 6.3 Results of the precipitation-runoff model; hydrographs for the predictive modeling period; the relevant period for the groundwater flow model is marked with the grey shaded area.

6.2 Groundwater Modeling Results Before Calibration

In this section the initial groundwater flow modeling results using uncalibrated model parameters are presented. Calculated water table levels were compared with measurements. A direct comparison of calculated values obtained from the uncalibrated model with observed values is illustrated in Figure 6.4. The straight line in the graph indicates a perfect fit of modeled values to measurements. Statistics and the summary of model performance criteria are summarized in Table 6.1.

Table 6.1 Statistics of the initial simulation

Criteria	Result
Residual Mean (m)	6.74
Res. Std. Dev. (RMSD) (m)	18.25
Sum of residual squares (m^2)	18925.48
Abs. Res. Mean (m)	12.59
Min. Residual (m)	-18.55
Max. Residual (m)	90.18
Range in Target Values (m)	326.6
Std. Dev./Range of observed values (%)	5.59

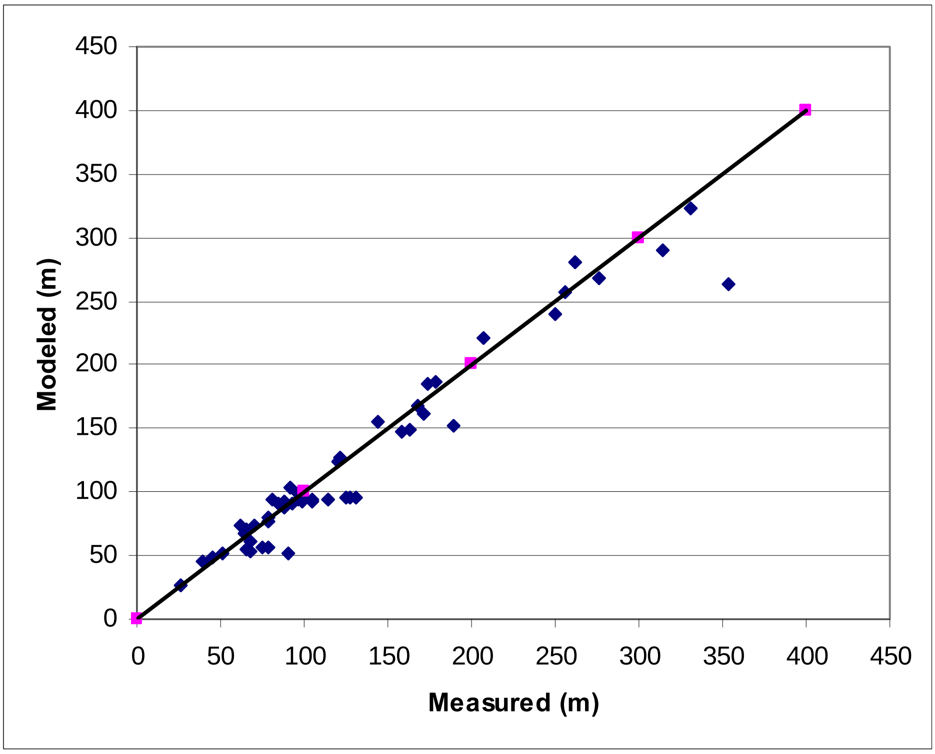

Figure 6.4 Comparison of modeled with measured head values for the uncalibrated model

6.3 Calibrated Groundwater Flow Modeling Results

Calibrated model parameters are presented in Tables 6.2 and 6.3. Calibrated hydraulic conductivity values remained within defined calibration boundaries except for zone 2, which represented the allochthonous limestone formation. Here the hydraulic conductivity value was decreased to the lowest possible value of 0.01 m/d. The recharge rate for zone 3 was decreased to the lower bound of the calibration range.

Model performance criteria and calibration statistics are given in Table 6.4. Overall, the calibrated groundwater flow model yielded satisfactory calibration statistics; residuals were distributed randomly around zero (Figure 6.5) and the residual mean, the absolute residual and the root mean squared residual (RMSD) were determined as 0.6, 11.0 and 16.4 m, respectively. The RMSD value was only 5% of the range of measured values. Overall, these values were acceptable within predefined model performance limits.

Table 6.2 Calibrated values for the hydraulic conductivity parameter

Zone	Calibration Interval	Initial Value	Calibrated Value
	Hydraulic Conductivity (m/d)		
K_{xy}-1 (Alluvial)	1 ~ 100	30	7.06
K_{xy}-2 (Allochthonous Limestone	0.01~10	0.03	0.01
K_{xy}-3 (Flysh)	0.001 ~ 1	0.3	0.30
K_{xy}-4 (Tuff)	0.1 ~ 10	0.3	7.09
K_{xy}-5 (Conglomerate)	0.05 ~ 5	1	1.91
K_{xy}-6 (Clayey Limestone)	0.05 ~ 5	1	1.35

Table 6.3 Calibrated values for the recharge parameter

Parameter	Calibration Interval	Initial Value	Calibrated Value
	Recharge (m/d)		
Zone 1- R1	$5\times10^{-4} \sim 1\times10^{-6}$	3×10^{-5}	6.27×10^{-5}
Zone 2- R2	$5\times10^{-4} \sim 1\times10^{-6}$	7×10^{-5}	1.27×10^{-4}
Zone 3- R3	$5\times10^{-4} \sim 1\times10^{-6}$	1.5×10^{-4}	0.0005

Model performance criteria are project-specific and no universal criteria exists. However, there are certain guidelines to obtain a successfully calibrate a groundwater flow model. The guidelines published by ASTM (2008) were taken as basis to evaluate the model performance for this study. Furthermore, it is evident from Figure 6.5 that the model yielded comparable water table elevations for most of the observation points; however, it was less successful for points located at higher elevations of the study site. The linear correlation coefficient (Zheng & Bennett, 2002) was calculated as 0.977, which indicates positively correlated observed and modeled head values. Better calibrated models tend to have linear regression coefficients close to 1.

Table 6.4 Statistics for the uncalibrated and the final calibrated model

Criteria	Uncalibrated	Calibrated model
Residual Mean (m)	6.74	0.63
Res. Std. Dev. (RMSD) (m)	18.25	16.40
Sum of residual squares (m^2)	18925.48	13476.83
Abs. Res. Mean (m)	12.59	11.00
Min. Residual (m)	-18.55	-36,44
Max. Residual (m)	90.18	74.51
Range in observed values (m)	326.6	326.6
Std. Dev./Range of observed values (%)	5.59	5.02

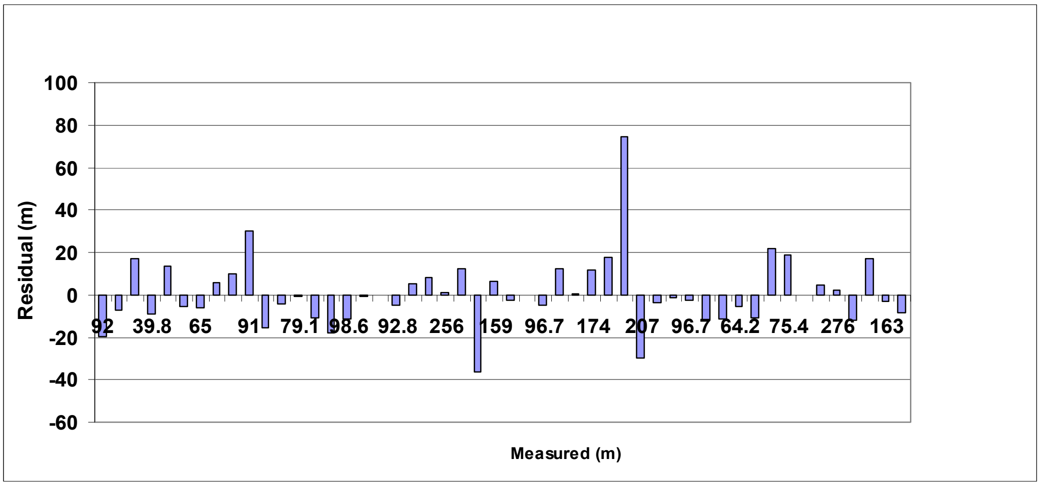

Figure 6.5 Comparison of residuals with observed values for the calibrated model

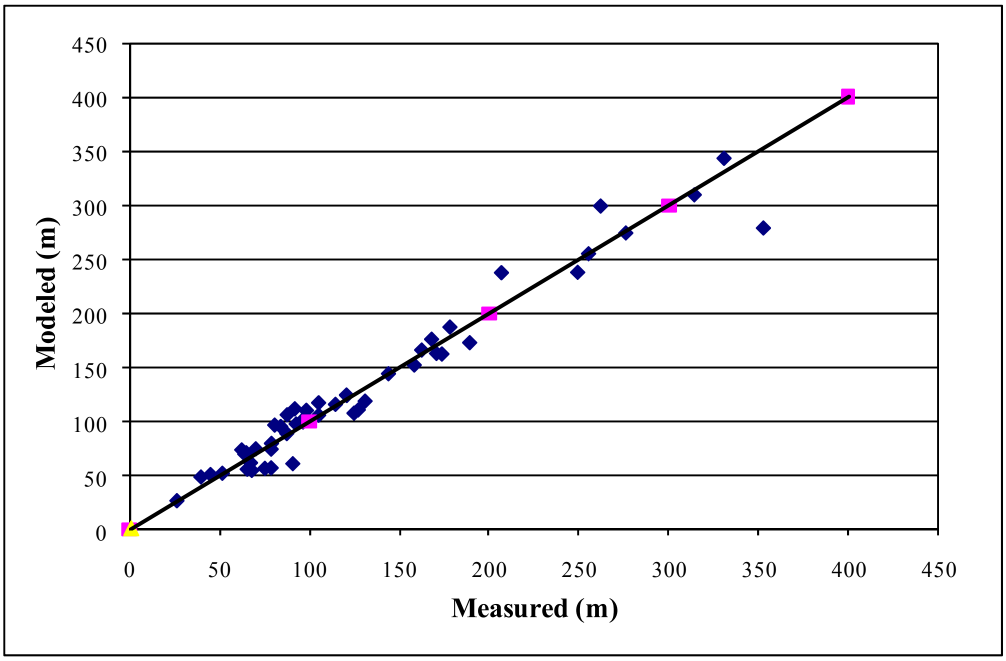

Figure 6.6 Comparison of modeled with measured head values for the calibrated model

Calculated water table elevations and groundwater flow directions are shown in Figure 6.7. These were reasonable and consistent with field observations. Based on the modeling results a strong influx of groundwater toward the watershed was observed at the foothills of the Nif mountain, in the northeast of the study area. The flow directions around the reservoir varied significantly, influenced mainly by stream-aquifer interactions and groundwater withdrawals. Water budget results of the model, shown in Table 6.5, revealed that groundwater recharge comprised about 20% of the total water input for the entire study area. Recharge was the second largest component in the budget after leakage from streams into the subsurface. It is also observed that there is a significant amount of ground water influx to the Tahtalı reservoir, when ground water flow directions in the vicinity of the reservoir are examined.

Table 6.5 Water budget of the calibrated model (winter conditions)

Flow rate (m^3/d)		**Flow rate (m^3/d)**	
IN		**OUT**	
Constant Head	121.90	Constant Head	123196.57
Wells	0	Wells	25859.40
Streams	290320.21	Streams	196164.95
Head dependent boundaries	32283.50	Head dependent boundaries	55435.96
Recharge	83066.98	Recharge	0
Total In	**405792.62**	**Total Out**	**405789.96**

Examining the water table contour map (Figure 6.7) reveals other interesting results; a groundwater mound is formed near Gaziemir, where flow diverges in several directions. It can be also noted that a hydraulic connection exists between the Nif mountain and the Tahtalı dam reservoir, although this needs to be confirmed with particle tracking simulations studies. Based on the model water budget, the total inflow to the Tahtalı reservoir is 100456 m^3/d and the outflow is only 116 m^3/d. This means that the reservoir gains water from the surrounding aquifer. Furthermore, the hydraulic gradient in the urban part of the study area (north) is relatively steep, in particular in the northwest, where elevated groundwater flow velocities are expected to occur. This result can be confirmed with the rough topography and steep terrain in that region. Flow in the Cumaovası alluvial basin is generally to the south, towards the Tahtalı reservoir.

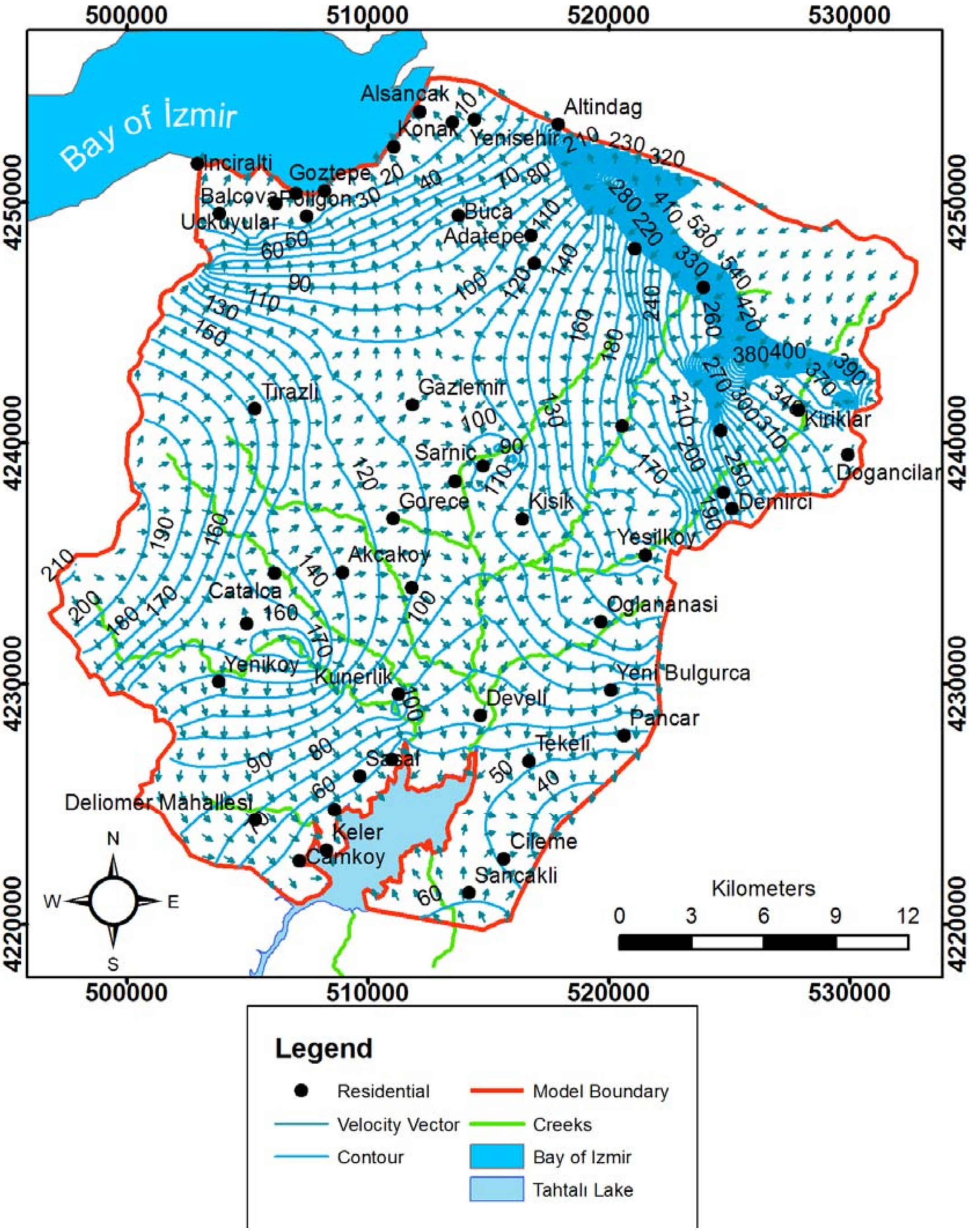

Figure 6.7 Groundwater flow modeling results: water table elevation contours and groundwater flow directions

Furthermore, groundwater flow interchange between the neighboring Torbalı sub-basin in the southeast of the study area can be observed. According to the modeling results, a net groundwater flux of 31390 m^3 per day is estimated to occur from the Tahtalı stream basin to the Torbalı basin.

6.4 Model Verification

The model was verified against the October head measurements, which were representative of dry summer conditions. Groundwater levels at 43 wells were October 2007 season, which were used as targets in the verification process. Boundary conditions and recharge rates were modified accordingly to match seasonal conditions of the study area. The model was run once using the modified model parameters and the steady-state simulation for summer conditions was obtained.

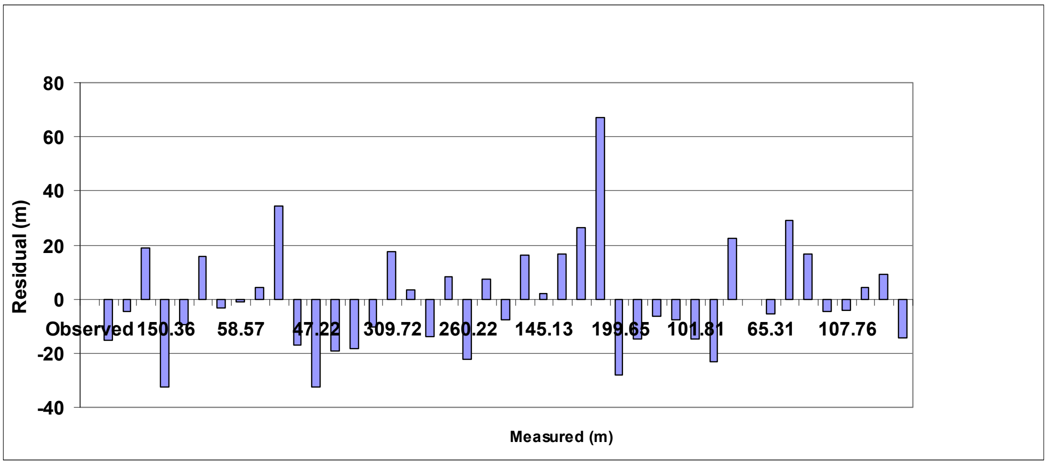

Figure 6.8 Comparison of residuals with observed values for the verified model

Model performance statistics pertaining to the verification run are given in Table 6.6. The model performed well but statistics indicated poorer performance compared to the calibrated model. Nevertheless, most criteria were within acceptable limits. Residuals were again distributed randomly around zero (Figure 6.8) and the residual mean, the absolute residual and the root mean squared residual (RMSD) were determined as -0.27, 15.2 and 19.4 m, respectively. The RMSD value was 6.3% of the range of measured values.

Table 6.6 Model performance statistics of the verification run

Criteria	Result
Residual Mean (m)	-0.27
Res. Std. Dev. (RMSD) (m)	19.40
Sum of residual squares (m^2)	16194.07
Abs. Res. Mean (m)	15.15
Min. Residual (m)	-32.63
Max. Residual (m)	67.04
Range in Target Values (m)	308.49
Std. Dev./Range of observed values (%)	6.29

Shown in Figure 6.9 is the comparison of modeled against measured hydraulic head values. Systematic errors are not visible in this scattergram. The linear correlation coefficient was calculated as 0.969, slightly lower than the coefficient for the calibrated model. The water budget for the verified model is summarized in Table 6.7. In comparison to the calibrated model, total groundwater recharge decreased in the summer 12.5% to 72683.61 m^3/d. Net groundwater flux from the Tahtalı basin to the neighboring Torbalı basin increased slightly to 34801 m^3 per day.

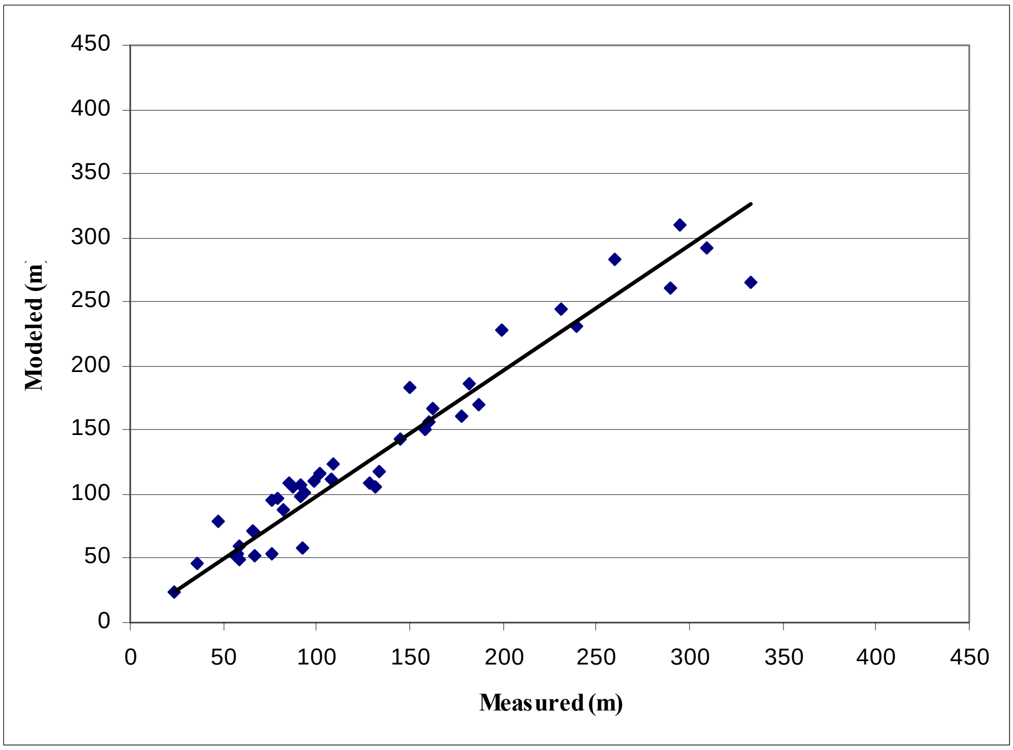

Figure 6.9 Comparison of modeled with measured head values for the verified model

Table 6.7 Water budget for the verified model (summer conditions)

Flow rate (m^3/d)		Flow rate (m^3/d)	
IN		**OUT**	
Constant Head	1345.69	Constant Head	131275.60
Wells	0	Wells	25859.40
Streams	300504.03	Streams	189738.60
Head dependent boundaries	32928.94	Head dependent boundaries	56370.14
Recharge	72683.61	Recharge	0
Total In	**407462.25**	**Total Out**	**405789.97**

6.5 Seasonal Decline of the Water Table

The seasonal decline of the water table was evaluated by obtaining the difference between the wet (calibrated model) and dry period (verified model) water table elevations. Shown in Figure 6.10 is the spatial distribution of the water table decline. Groundwater level decline values exhibit some variation both spatially and quantitatively. Based on this map, it is evident that the groundwater decline was mostly less than 1 m. Decline was calculated up to 3 m in the east and west of the Tahtalı reservoir. This is partly due the fact that the geology is significantly different in these areas. Furthermore, increased groundwater withdrawal during the summer season is also an important factor. In the alluvial basin of the Cumaovası plain, decline is limited due to relatively higher hydraulic conductivities in the aquifer, despite increased water withdrawal in the summer. On the other hand, maximum decline values were observed in the northeast of the study area, in the vicinity of the Nif Mountain. Here the hydraulic conductivities are lower compared to other areas. This implies that lower groundwater recharge translates to much lower water table elevations.

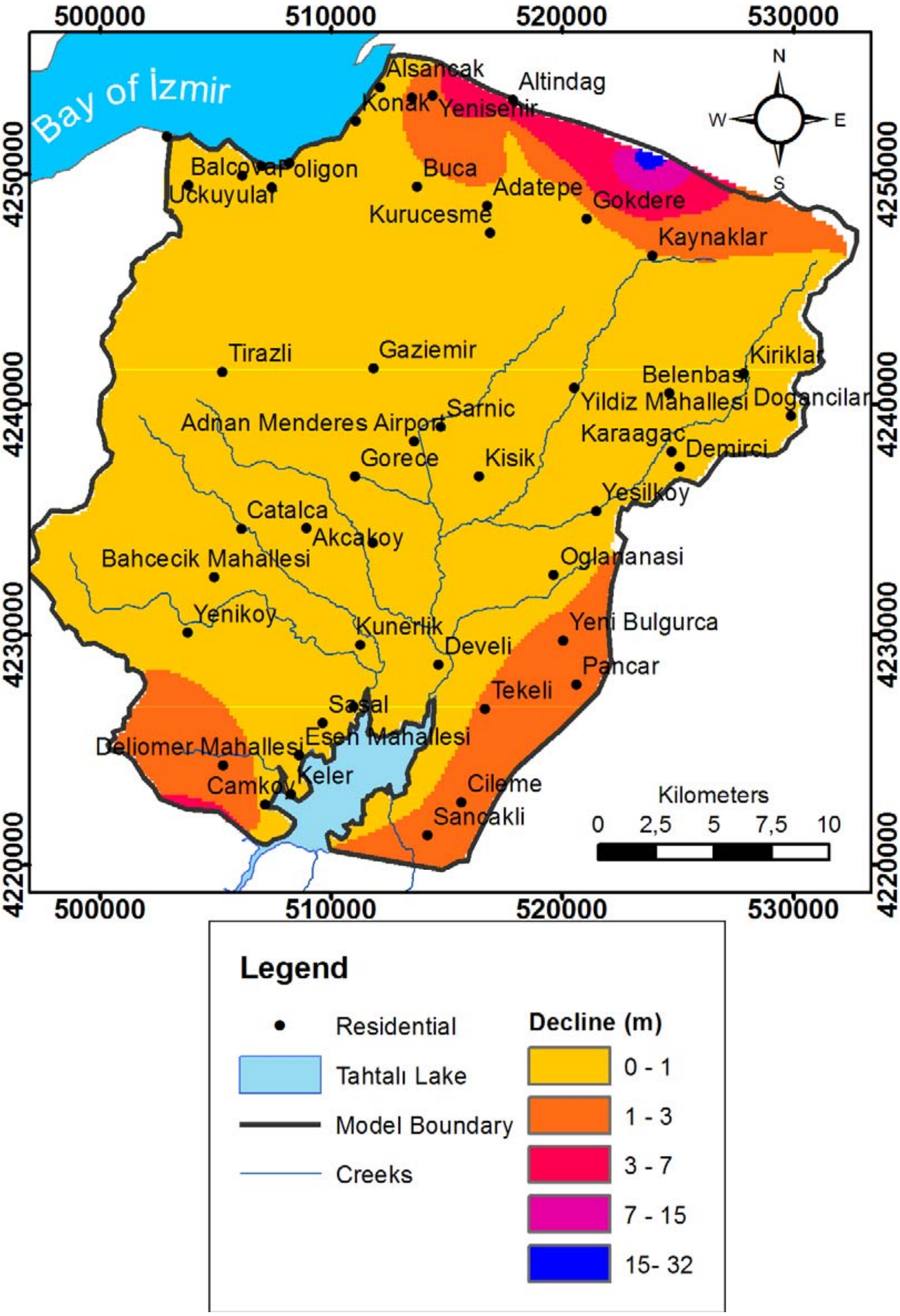

Figure 6.10 Calculated seasonal decline of the water table

CHAPTER SEVEN
DISCUSSION AND CONCLUSIONS

7.1 General Discussion and Shortcomings

A modeling study of groundwater flow for the southern part of the metropolitan area of Izmir, which entirely includes the important Tahtalı stream basin, was presented. A comprehensive model was developed with a combined precipitation-runoff modeling to estimate groundwater recharge. Water table elevations, groundwater flow directions, seasonal decline in groundwater levels, recharge rates, flux to the neighboring Torbalı sub-basin and water budgets were determined with the developed model for the dry summer and wet winter seasons. Water budget results of the model revealed that groundwater recharge comprised about 20% of the total water input for the entire study area. Recharge was the second largest component in the budget after leakage from streams into the subsurface. The modeling results also hint at the likelihood that the water quality of the Tahtalı reservoir is probably influenced by activities in the rural part of the study area, e.g. the Cumaovası plain. However, to better evaluate the vulnerability of water resources in the area to diffuse pollution, a contaminant transport modeling study that is based on the presented flow model may be warranted. Therefore, contaminant transport model could be run within the basin boundary.

Furthermore, it was demonstrated with this study that a robust modeling approach can be taken by combining results of a lumped, water budget based precipitation-runoff model with a distributed groundwater flow model. Groundwater recharge in groundwater flow models is often one of the most uncertain model parameters since it is almost impossible to measure it directly in the field for large watersheds. Nevertheless, it is important to somehow quantify recharge, in particular for diffused pollution vulnerability studies.

Some shortcomings and limitations of the developed groundwater flow model are discussed in the following section:

1. It is conceivable that the well measurements did not reflect the true depth to the water table due to the fact that the monitoring wells were actually irrigation wells with very long well screens; in some cases longer than 100 m. It is likely that this fact affected the performance of the model.
2. Related to the previous item; the monitoring wells were screened over several aquifer units and sometimes over units with different properties. Therefore, the representativeness of the well measurements is somewhat questionable. Nevertheless, conceptually the groundwater flow model would not be different if perfect measurements would be available, only the accuracy of the model would be better.
3. Errors in SRTM data are likely to have affected the calibration of the model because they were used to determine water table elevations.
4. Monitoring wells are sparse in some parts of the study area. Accessible monitoring wells were unavailable in particular in mountainous parts of the study area or in areas where groundwater was either deep or not available. A denser network of wells around the Tahtalı reservoir and more wells in the urban area would have increased the quality of the model.
5. The precipitation-runoff model provided a recharge rate that is uniform over a relatively large area. This is due to the fact the precipitation-runoff model was not a distributed model and was based on water budget calculations that were only valid for the entire watershed. Dividing the recharge zone that represented the Tahtalı stream watershed into smaller sub-watersheds would have enhanced the outcome of the groundwater flow model. However, this was not possible because absent observation data for the sub-watersheds in the study area.
6. The amount of groundwater withdrawal in the study area could only be grossly estimated. The actual amount is unknown and hard to quantify since numerous irrigation wells exist in the fertile plains. Many wells are not licensed and are not accounted for by the water authorities. Therefore, the total groundwater withdrawal is expected to be much higher. It is

possible to enhance the groundwater flow model through more additions of pumping wells in the study area.

7.2 Recommendation for Future Work

The presented groundwater flow model and the precipitation-runoff model can be undoubtedly improved. Also, the purpose and thereby the application of the model can be re-defined. Recommendations for future studies can be listed as follows:

1. Recharge rates used in the model can be modified to accommodate climate change scenarios to eventually assess the effects of climate change on water resources in the study area
2. Investigate more monitoring wells to improve the calibration of the model
3. Obtain more sets of monitoring data to improve the overall reliability and usability of the model
4. Inclusion of more pumping wells to account for a more accurate groundwater withdrawal
5. Enhance the calibration of the groundwater flow model by including the calibration to springflow measurements
6. Revisit the parameters and formulations of the precipitation-runoff model for higher model accuracy
7. Conduct particle-tracking simulation to support the interpretation of modeling results
8. The model results can be used as input for contaminant transport modeling studies in order to evaluate the effects of different land-use practices or diffuse pollution scenarios.

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