SIMULATION OF GROUNDWATER FLOW IN CACHE VALLEY, UTAH AND IDAHO

by

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ABSTRACT

Simulation of Groundwater Flow in Cache Valley,

Utah and Idaho

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A groundwater model of Cache Valley was created using MODFLOW. Steady-state calibration of the model demonstrated that recharge to the lower confined aquifer may occur along the margin of the valley that borders the Wellsville Mountains and the Bear River Range. Steady-state calibration also showed that discharge from the unconfined aquifer may occur along the eastern and western margins of the valley in both the Utah and the Idaho portions of the valley.

Two simulations were run with increased pumping of 35 cubic feet per second (1 cubic meter per second) from the principal aquifer. The first simulation was run with the average annual precipitation value of 1.2 feet per year (0.36 meters per year), while the second was run with a less than average annual precipitation value of 1 foot per year (0.3 meters per year).

The first simulation produced very little change within the unconfined aquifer. The discharge from the groundwater system through springs, seepage to streams, evapotranspiration, and general head boundaries remained unchanged with the increase in discharge through pumping. This indicates that the two continuous, confining layers that blanket the valley may serve as a barrier to groundwater flow between the unconfined and lower confined aquifer. The increased pumping within the principal aquifer did not stimulate increased recharge along the western margin of the valley. This indicates that true steady-state conditions were not achieved in the amount of time that the model had indicated.

During the second simulation, decreased recharge to the groundwater system through infiltration of precipitation caused a decrease in discharge from the groundwater system through seepage to streams, springs, evapotranspiration, and general head boundaries. The increased pumping within the principal aquifer also did not stimulate increased recharge along the western margin of the valley. As with the first simulation, this indicates that true steady-state conditions were not achieved in the amount of time that the model had indicated.

A sensitivity analysis of the model concluded that the hydraulic conductivity of the two continuous, confining layers that blanket the valley proved to have a relatively substantial impact on the water levels in the confined aquifers. The sensitivity analysis also showed that altering the vertical hydraulic conductivity of the lower confined aquifer produced minimal head changes.

(89 pages)

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I am grateful to my parents, Sherman and Janyce Myers, for all their love and support. My parents always stressed the importance of hard work and dedication in accomplishing the things that I desired. This belief has greatly influenced my life.

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Barry Myers

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CHAPTER I

INTRODUCTION

Statement of the Problem

Rapid urbanization within many of Utah's river basins occurred in the latter part of the 20th century, including Cache Valley. This trend brings with it changes in water demands and increasing water quality problems. In order to protect surface water rights holders, as well as groundwater from long-term degradation, planning for the most efficient and equitable use of the water is important.

The framework for water resources management in Cache Valley has not been designed to rapidly adjust to the changing needs that urbanization is placing on different water agencies and stakeholders. The various agencies and stakeholders are fragmented in meeting their groundwater objectives. They are not taking advantage of the common ground that exists among them, such as the overlap of the various users' information needs.

With increasing urbanization of Cache Valley, the necessary planning to avoid future problems is becoming increasingly complex. The decisions made now will dictate whether future problems will be reduced or exacerbated. In order to minimize future problems, a computer simulation of the groundwater resources in Cache Valley, using an accurate conceptual hydrostratigraphic model, is critical in aiding the decision making.

Objectives

The purpose of this project was to develop useful information about the groundwater resources of Cache Valley that can be used in future decision making. This was accomplished through the creation of a new MODFLOW (McDonald and Harbaugh 1988) model of Cache Valley. The final product is a groundwater model of Cache Valley that is calibrated to 1999 steady-state conditions. Two simulations were executed using this model. Through the simulation of various hydrologic conditions, the relative importance of obtaining accurate estimates of the various hydrologic parameters of the groundwater system can be assessed. The model indicates which parameters are in need of further monitoring, and which parameters are less likely to require long-term monitoring. If used properly, this model can be a valuable management tool for resource planners.

Location

Cache Valley is a north-south trending basin. Slightly more than the southern half of the valley lies in Utah, with the northern portion lying in Idaho (Figure 1). The Bear River Range comprises the eastern boundary of the valley. The Bannock Range, Malad Range, and the Wellsville Mountains provide the valley with its western boundary. The valley is approximately 70 miles (110 kilometers) long, 16 miles (26 kilometers) wide at its widest point, and has an area of 660 square miles (1700 square kilometers). The valley lies entirely within the Bear River drainage basin. The Bear River enters the valley near its northern margin and exits the valley to the west through Cutler Narrows.

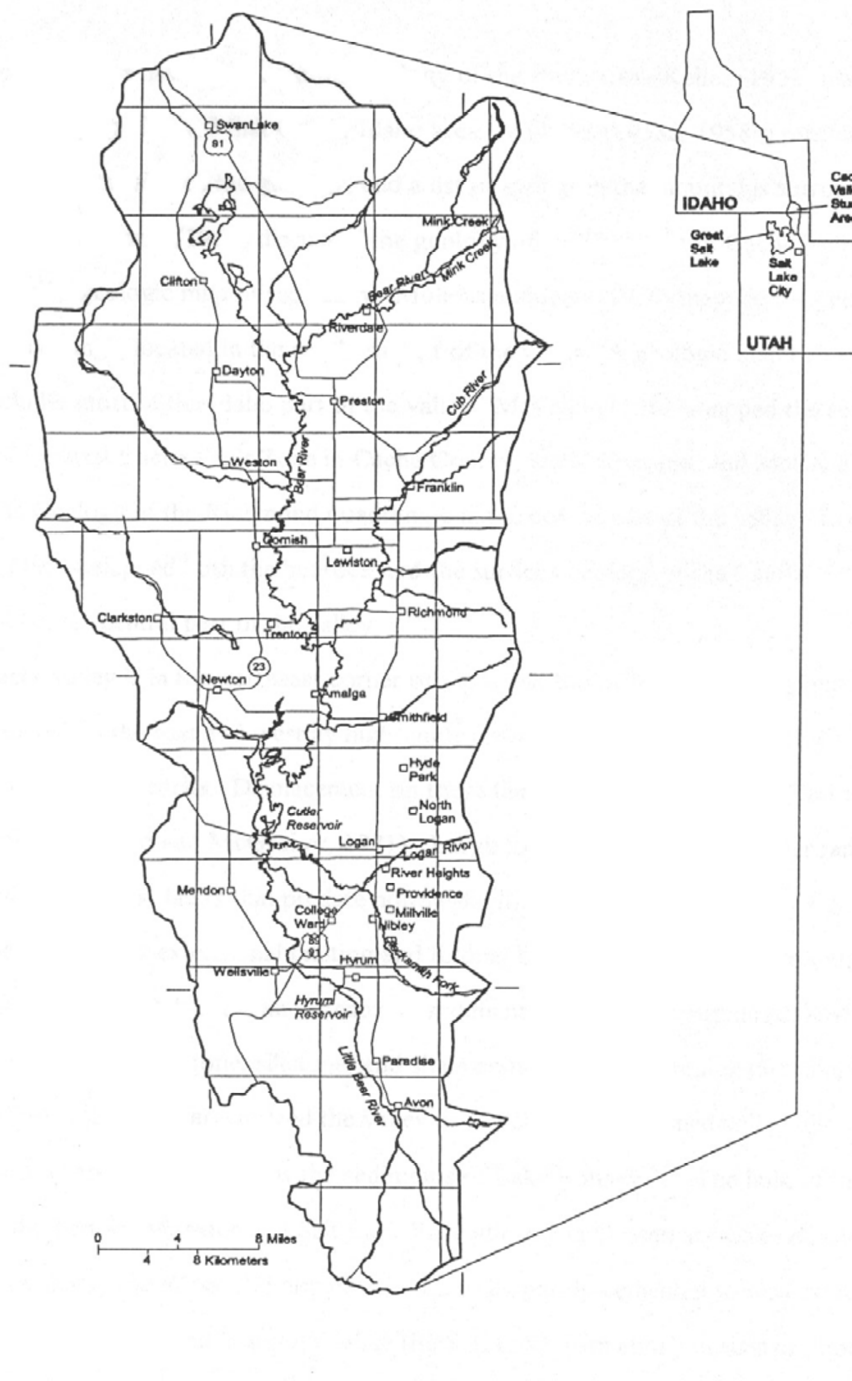

Figure 1: Location of Cache Valley study area (Robinson 1999).

CHAPTER II

BACKGROUND

Geologic Setting

Cache Valley is located in the northeastern corner of the Basin and Range Province. It is a narrow, elongate, and complex graben at the southwest end of a series of half-grabens that form an extensional corridor between the Wasatch and Teton normal faults (Evans and Oaks 1996). Evans and Oaks (1996) also maintain that the basin is a narrow, deep half-graben above a single west-dipping, listric normal fault at its southern end. In the central part, the basin is a doubly tilted graben bounded on both sides by normal faults. The basin is broad, shallow, and flat-bottomed at its northern end. The valley's eastern edge is bounded by the East Cache normal-fault zone, and the western margin is bounded by the West Cache fault zone (McCalpin 1989). The East Cache fault zone is listric, and the West Cache fault dies out or splits into a series of poorly exposed splays southward (Evans and Oaks 1996).

The footwalls on either side of the valley are composed of Proterozoic and Paleozoic rocks which include limestone, dolostone, sandstone, and shale. These rocks also underlie the younger, unconsolidated sediments of the basin (Williams 1962). Five north-south and east-west seismic-reflection profiles across the valley examined by Evans and Oaks (1996) suggest that basin width, sedimentary basin-fill thickness and geometry, fault geometry, and fault slip vary from north to south.

The Salt Lake and Wasatch Formations are the two units in the valley that are of Tertiary age. The Cenozoic Wasatch Formation is a poorly cemented to well-cemented conglomerate and sandstone (Bjorklund and McGreevy 1971). Evans and Oaks (1996) infer that the Wasatch Formation was continuous across the study area before the onset of extension. Oaks and Runnells (1992) found that the Wasatch Formation is at least 803 feet (245 meters) thick in the Bear River Range and 328 feet (100 meters) thick in the central part of the basin.

The Salt Lake Formation overlies the Wasatch Formation. Williams (1962) suggests that the Salt Lake Formation consists of conglomerates, tuffaceous sandstones and siltstones, and limestones, and that it is exposed in an almost continuous belt in the foothills around the valley. Evans and Oaks (1996) maintain that the formation is thickest along the eastern margin of the valley, at approximately 9,000 feet (2,700 meters), and thins toward the west.

Robinson (1999) concluded that the conglomerates of the Salt Lake Formation consist of sub-rounded to well-rounded coarse sand to boulders. The conglomerates are clast supported and often have a tuffaceous, sandy groundmass. The tuffaceous unit consists of a light tan to olive gray, tuffaceous claystone with beds and lenses of gray volcanic ash.

In the valley interior, at least several hundred feet of fluvial and lacustrine sediments of Quaternary age underlie Lake Bonneville deposits and overlie the Salt Lake Formation (Williams 1962). Through the interpretation of well logs, Robinson (1999) found that in the center of the valley these sediments are more than 800 feet (240 meters) thick. Williams (1962) maintains that the contact of these sediments with the overlying lake bottom deposits is irregular, but generally lies 60 to 150 feet (18 to 46 meters) below the ground surface.

The Little Valley lake cycle is the name given to the penultimate cycle of Lake Bonneville (Scott et al. 1983). This cycle occurred between 90,000 and 150,000 years ago. Scott et al. (1983) found the highest level of this lake cycle to be 246 to 393 feet (75 to 120 meters) below the Bonneville shoreline. The deposits of this lake cycle consist primarily of near-shore sand and gravel and shallow-water deposits of marly silt and fine sand. Lakes of this cycle probably spilled over into Cache Valley, but the lake cycle's sediment record has yet to be confirmed in Cache Valley.

During the episode between the Little Valley lake cycle and the start of the Bonneville lake cycle, which began 25,000 to 30,000 years ago, the deposits of Little Valley age were eroded, incised, and in places, buried by sub-aerial sediments (Scott et al. 1983).

The Lake Bonneville lake cycle has had the most influence in shaping the landscape that exists in Cache Valley. Between 25,000 and 30,000 years ago, the lake rose steadily, with pauses that lasted as long as 1,500 years, and reached the Bonneville shoreline in areas of low isostatic depression about 16,000 years ago (Scott et al. 1983). The lake level remained close to the Bonneville shoreline until approximately 15,000 years ago, after which the lake level dropped 328 feet (100 meters) during the Bonneville flood. Subsequently, the Provo shoreline formed. This shoreline forms Cache Valley's prominent shoreline and deltaic features. By 13,000 years ago, the lake had fallen below the Provo level, and Lake Bonneville retreated out of Cache Valley for the last time. By 11,000 years ago, the lake stood close to the level of present Great Salt Lake.

Sediments deposited by Lake Bonneville include the Alpine and Bonneville Formations, which consist mostly of silt with some gravel, and the overlying Provo Formation, which consists of interbedded layers of gravel, sand, silt, and clay (Williams 1962). Gravel and sand of Lake Bonneville age were deposited as shore embankments, deltas, bars, and spits near the mountain fronts, while silt and clay settled from suspension in the lake water at lower altitudes in Cache Valley.

Previous Hydrogeological Investigations

Through the compilation of geohydrologic sections, Bjorklund and McGreevy (1971) estimated the fill deposits of Cache Valley to be more than 5,000 feet (1,500 meters) of gravel, sand, silt, clay, and conglomerate of Tertiary and Quaternary age, and to contain more than 40 million acre-feet (50 billion cubic meters) of groundwater. Much of the groundwater is found within fine-grained deposits, which do not readily yield water, rendering them unsuitable for groundwater development. The groundwater in the fill is found in both confined and unconfined systems. The compilation of geohydrologic sections and the analysis of well logs were used to construct a hydrostratigraphic conceptual model of the groundwater system. This conceptual model is shown in Figure 2(A). This model depicts one thick clay layer that is continuous throughout the areal extent of Cache Valley, and terminates approximately one mile (1.6 kilometers) from the valley margin.

In the Smithfield-Hyrum-Wellsville area, Bjorklund and McGreevy (1971) report that the yields to wells are as much as 8 cubic feet per second (0.2 cubic meters

Figure 2: Hydrogeologic conceptual model developed by (A) Bjorklund and McGreevy (1971) and (B) Kariya et al. (1994).

per second), and the transmissivity is as much as 330,000 square feet per day (31,000 square meters per day). This is the largest and most productive aquifer system in the valley. This aquifer is regarded as an overflowing groundwater system. The second most productive aquifer system is found on the west and north margin of the valley.

Bjorklund and McGreevy (1971) maintained that groundwater levels have fluctuated with fluctuations in annual recharge, but that overall, the groundwater levels in the main aquifer in Cache Valley had not changed from 1935 until the time of their study.

Along the east side of Cache Valley, groundwater moves from the recharge areas along the mountain front of the Bear River Range westward toward the lower parts of the valley. Groundwater from the front of the Bannock Range flows east and southeast toward discharge areas along the Bear River. Groundwater is discharged from bedrock to the valley fill in the southwestern portion of the valley. Within the groundwater flow regime of Cache Valley, some groundwater systems are hydraulically connected, while others are considered separate systems. Bjorklund and McGreevy (1971) also found that 4,000 acre-feet (5 million cubic meters) of groundwater moves annually from Idaho into Utah in Cache Valley. They estimated this to be 3,000 acre-feet (3,700,000 cubic meters) in the area west of the Bear River near Weston, Idaho and Cornish, Utah, and approximately 1,000 acre-feet (1,200,000 cubic meters) in the Cub River subvalley mostly east of the Cub River.

Bjorklund and McGreevy (1971) found that recharge to the principal groundwater system in Cache Valley occurs mainly through infiltration of water from precipitation, streams, canals, ditches, irrigated fields, and by subsurface inflow. The principal recharge areas are along the margins of the valley, with some recharge to shallow, unconfined aquifers in the lower portions of the valley.

Much of the groundwater reservoir was overflowing. Bjorklund and McGreevy (1971) therefore assumed that the change in storage was negligible. They regarded the total recharge to be about equal to the total discharge. They calculated about 280,000 acre-feet (350,000,000 cubic meters) of annual recharge on this basis.

Bjorklund and McGreevy (1971) also generated a water-budget analysis for Cache Valley. This is summarized in Table 1. Their water budget analysis indicates that an average amount of about 2,350 cubic feet per second (67 cubic meters per second) of water enters and leaves the valley annually, and that annual changes in both surface water and groundwater storage are considered to be negligible.

Kariya et al. (1994) conducted a hydrologic study of Cache Valley using hydrologic data collected by the U.S. Geological Survey during 1989-1992. Through this study, a groundwater budget was estimated (Table 2). This budget represents recharge and discharge of water to the main groundwater system in the unconsolidated basin-fill deposits of Cache Valley. Some of the groundwater budget components were estimated as part of the previous study of Cache Valley by Bjorklund and McGreevy (1971), while other budget components were modified. Other forms of recharge include subsurface inflow from adjacent consolidated rock and unconsolidated basin-fill deposit groundwater systems, and seepage from ephemeral streams. This number is the difference between total discharge and

Table 1: Total Water Budget Analysis by Bjorklund and McGreevy (1971).

Inflow	**Flow (cubic feet per second)**
Principal streams, average 1960-68 water years	1380
Runoff not included in principal streams	40
Imports through pipelines for public supply	35
Springs near valley edge	55
Precipitation on Cache Valley	804
Groundwater inflow not accounted for in springs near the valley edge	44
Total (rounded)	2350
Outflow	
Streams, average 1960-68 water years	1390
Consumptive use by phreatophytes	149
Consumptive use on irrigated land	463
Consumptive use in urban areas	58
Consumptive use on dry farm and noncleared land	268
Evaporation from open water	28
Groundwater outflow	Negligible
Total (rounded)	2350
Change in surface-water storage	Negligible
Change in groundwater storage	Negligible

Table 2: Groundwater Budget for Cache Valley (Kariya et al. 1994). Some of the flow rates listed under Bjorklund and McGreevy (1971) were modified by Kariya et al. (1994) and are not consistent with the values reported by Bjorklund and McGreevy (1971).

	Bjorklund and McGreevy (1971) (cubic feet per second)	**Kariya et al. (1994) (cubic feet per second)**
Recharge		
Infiltration from precipitation and unconsumed irrigation water	186	57
Seepage from canals	160	140
Seepage from streams	7	3
Other forms of recharge	96	96
Total Recharge	449	296
Discharge		
Seepage to streams and Cutler Reservoir	180	180
Spring discharge	138	138
Evapotranspiration	87	87
Withdrawal from wells	44	52
Total Discharge	449	457

recharge from infiltration of precipitation and unconsumed irrigation water and seepage from canals and streams.

Kariya et al. (1994) also created a hydrostratigraphic conceptual model of the unconsolidated basin-fill deposits of Cache Valley (Figure 2(B)). This model was constructed through the use of various geohydrologic sections of Cache Valley compiled by Bjorklund and McGreevy (1971). This conceptual model indicates that discontinuous layers of clay occur in most of the unconsolidated basin-fill deposits, but it lacks the continuous confining layer that was described by Bjorklund and McGreevy (1971).

As part of the study, Kariya et al. (1994) used MODFLOW (McDonald and Harbaugh 1988), a modular, three-dimensional, finite-difference, groundwater-flow model, to simulate flow in the groundwater system in the unconsolidated basin-fill deposits of Cache Valley. A rectangular grid composed of 82 rows and 39 columns represented the unconsolidated basin-fill deposits. The basin-fill deposits were also represented with six model layers to enable simulation of vertical gradients. The model simulated confined and unconfined conditions, withdrawal from wells, areal recharge, evapotranspiration, seepage to drains, and seepage to and from streams and consolidated rock (Kariya et al. 1994).

The model was calibrated to 1969 steady-state and 1982-1990 transient-state conditions. Results of the steady-state calibration were used as initial conditions for the 1982-1990 transient-state calibration (Kariya et al. 1994). Bjorklund and McGreevy (1971) found no change in storage or long-term water levels, therefore, steady-state conditions were assumed for 1969. The transient-state calibration was done by simulating the groundwater system using 1-year stress periods for 1982-1990 (Kariya et al. 1994).

During transient-state calibration, Kariya et al. (1994) discovered that recharge from precipitation and unconsumed irrigation water did not vary sufficiently from one year to the next to reproduce the observed changes in water levels. The initial estimates of recharge from precipitation and unconsumed irrigation water were altered using a Deep Percolation Model (Bauer and Vaccaro 1987).

Through a comparison of model-computed and estimated fluxes, Kariya et al. (1994) indicated that 104 cubic feet per second (2.9 cubic meters per second) of water was lost from storage in 1990. During the 1982-1990 time period, water levels fell and groundwater was lost from storage during 1982-1984 and in 1986. During the remaining years, groundwater went into storage.

After calibration, two simulations were run. When the dry conditions of 1990 were simulated for 5 years, water levels declined more than 20 feet (6 meters) in the south end of the valley. Water levels declined about 10 feet (3 meters) between Richmond and Hyrum (Kariya et al. 1994).

Kariya et al. (1994) also simulated increased pumpage by adding three well fields in the Logan, Smithfield, and College Ward areas. Each of these well fields pumped 10 cubic feet per second (0.3 cubic meters per second). After simulation, water level declines of as much as 51 feet (16 meters) were projected in areas close to the well fields, and declines greater than 10 feet (3 meters) were projected in most of the southeastern part of the valley. According to Kariya et al. (1994), these results suggest that the groundwater system should approach a new steady state after about 30 years.

Through their study, Kariya et al. (1994) found that discharge from the groundwater system in the unconsolidated basin-fill deposits includes seepage to streams and reservoirs, spring discharge, evapotranspiration, and withdrawal from wells. Kariya et al. (1994) also maintained that because of the interconnection of the surface water and groundwater systems in Cache Valley, increased withdrawal of groundwater could decrease the volume of groundwater discharge to the surface water system, and is therefore of concern to surface water users.

Through the detailed study of drillers' logs, Robinson (1999) developed the most complete hydrostratigraphic conceptual model of Cache Valley that has yet been developed. This new conceptual model (Figure 3) suggests that two continuous confining layers, aggregating to approximately 100 feet (30 meters) in total thickness, blanket the valley and terminate within about one mile (1.6 kilometers) of the valley margin. These clay layers correlate with the deposits of the Bonneville and Little Valley lake cycles, respectively. Below the confining layers, the Quaternary deposits have an aggregate thickness of more than 500 feet (150 meters). The Quaternary deposits near the eastern valley margin are composed of alluvial fan and deltaic sands and gravels. These deposits thin westward. The Quaternary sediments west of the alluvial fan and deltaic deposits are composed of well to poorly sorted sands and gravels, silts, and clays. Individual layers of these units are not continuous over large areas. According to Robinson (1999), these two types of Quaternary deposits comprise the principal aquifer system.

Figure 3: Hydrogeologic conceptual model developed by Robinson (1999).

Robinson (1999) found that the most important source of recharge to the principal aquifer system is water from streams, specifically the Little Bear River, Logan River, Blacksmith Fork, and Summit Creek. This conclusion was deduced from the evaporative signature of oxygen-18 and deuterium data. Robinson (1999) also suggested that seepage from canals upon the benches in the recharge zone may be largely responsible for facilitating the infiltration of water derived from these rivers. Furthermore, oxygen-18 and deuterium data suggest that precipitation onto the benches is not a major source of recharge to the principal aquifer system.

Through the use of tritium values, Robinson (1999) roughly delineated the extent of post-1952 recharged water. Given the distance of the recharge zone from the post-1952 water, Robinson (1999) estimated that groundwater is moving west through the aquifer at a rate of 0.96 to 1.8 feet per day (0.29 to 0.55 meters per day). Robinson (1999) also estimated that from 1952 to 1988, groundwater in the principal aquifer was replaced at an average rate of 41 cubic feet per second (1.2 cubic meters per second). This value corresponds almost exactly with the withdrawal rates from the wells in the principal aquifer. This suggests that withdrawal from wells alone is enough to account for all groundwater movement through the principal aquifer. Therefore, virtually no groundwater is seeping upwards from the principal aquifer through the confining layers covering Cache Valley.

If the Bear and other rivers in Cache Valley do not gain water from the principal aquifer, they must have some other source, as they are indisputably gaining streams (Robinson 1999). Robinson (1999) therefore suggests that these rivers are recharged through the shallow, unconfined aquifer.

Long-term measurements of water levels in the principal aquifer show no long-term declines. Robinson (1999) states that well pumping must somehow increase recharge to the aquifer. He found that the most likely mechanism to explain this increased recharge is a reduction in discharge from one or more of the many springs in Cache Valley. It appears that the springs in Cache Valley act as an overflow valve for the principal aquifer (Robinson 1999).

CHAPTER III

MODEL CREATION

MODFLOW (McDonald and Harbaugh, 1988), a modular, three-dimensional, finite-difference, groundwater-flow model, has been used to simulate flow in the groundwater system in the saturated, unconsolidated basin-fill deposits of Cache Valley. The model simulates confined and unconfined conditions, withdrawal from wells, areal recharge, evapotranspiration, and seepage to and from streams and unconsolidated rock.

Discretization of the Groundwater System

The portion of Cache Valley described by Bjorklund and McGreevy (1971) as the “valley floor” is considered to be the portion of the study area covered by saturated, unconsolidated basin-fill deposits. These saturated, unconsolidated basin-fill deposits have been discretized into a horizontal, rectangular grid composed of cells (Figure 4). The cells range in size from 1 mile by 1 mile to 0.5 miles by 0.375 miles (1.6 kilometers by 1.6 kilometers to 0.8 kilometers by 0.6 kilometers). This area is smaller than the drainage basin due to a large portion of the drainage basin being occupied by consolidated bedrock. The bedrock surrounding the valley and the unconsolidated, basin-fill deposits near the mountain fronts that overlie shallow bedrock either are not saturated or transmit very little groundwater, and therefore have not been simulated by the model. In the southeastern part of the basin, where there are more wells and a larger volume of groundwater withdrawal than in the northern portion, smaller cells have been used to provide finer resolution.

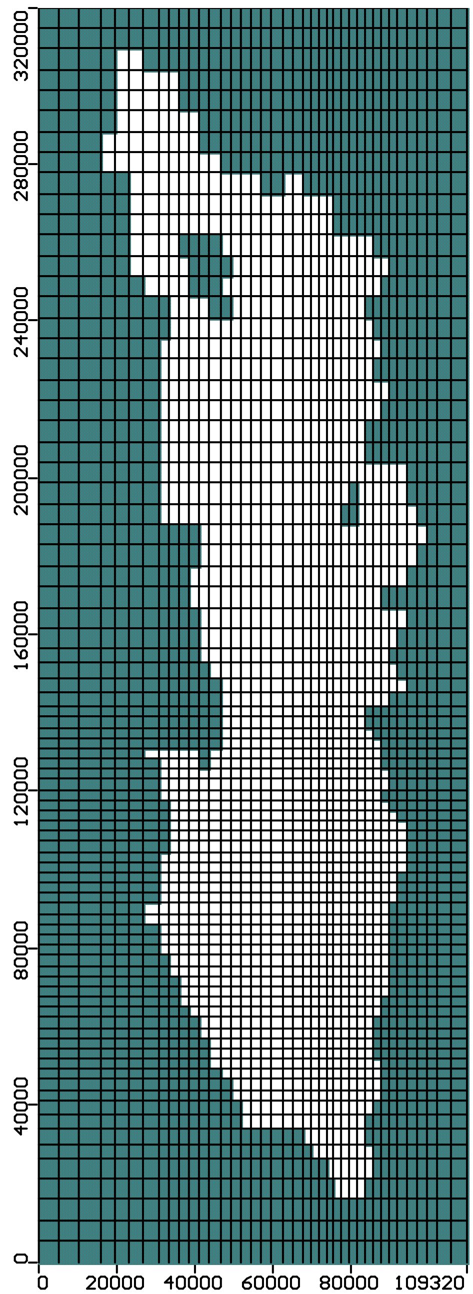

Figure 4: Rectangular grid representing the unconsolidated basin fill deposits. white cells represent active cells, while shaded cells represent inactive cells. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

The numerical model is based upon the conceptual model developed by Robinson (1999) (Figure 3). His conceptual model depicts an unconfined aquifer, an upper confined aquifer located between an upper and a lower confining unit, and a lower confined aquifer located below the lower confining unit. Eleven layers represent thesaturated, unconsolidated basin-fill deposits described by Robinson (1999). The thickness of each layer is uniform throughout the model. Figure 5 is a cross section of the model depicting the layers of the saturated, unconsolidated basin-fill deposits.

Layer 1 simulates the unconfined aquifer. Although the thickness of the unconfined aquifer varies, a thickness of 100 feet (30 meters) was chosen. This thickness was chosen due to various rivers within the study area cutting deeply into the unconsolidated deposits, which would cause cells in layer one to dry up during simulation if the layer was not thick enough.

Layers 2 and 4 simulate the upper and lower confining layers, respectively. Robinson (1999) describes the upper confining layer as having an average thickness of 60 feet (18 meters), while the lower confining layer has an average thickness of 30 feet (9 meters). These average thicknesses were used in the model to represent the thicknesses of the corresponding layers.

Robinson (1999) states that the upper confined aquifer has an average thickness of 30 feet (9 meters). Layer 3, which represents the upper confined aquifer, correspondingly has a thickness of 30 feet (9 meters).

Layers 5 through 11 represent the lower confined aquifer, which consists of the more than 1,000 feet (305 meters) of Quaternary deposits (Bjorklund and

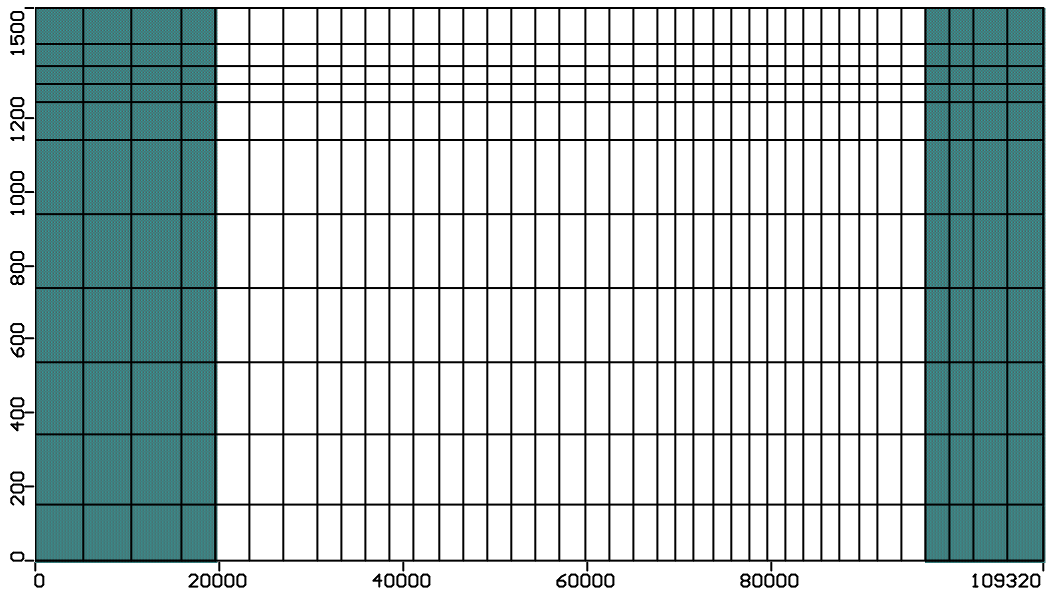

Figure 5: Cross section of the model depicting the various layers. The vertical scale is 40 times the horizontal scale. White cells represent active cells,while shaded cells represent inactive cells. The numbers on the vertical scale represent the depth of the model in units of feet, and the numbers on the horizontal axis represents the width of the model in units of feet.

McGreevy 1971). In order to reduce the risk of the model not converging to a solution, it is not recommended to drastically increase the size of adjacent cells. This was taken into account when selecting the thicknesses of layers 5 through 11. Layer 5 has a thickness of 100 feet (30 meters), Layers 6 through 10 have a thickness of 200 feet (61 meters), and Layer 11 has a thickness of 150 feet (46 meters).

Boundary Conditions and Groundwater Budget

The boundary conditions have been assigned as follows:

No-flow boundaries

- Between the unconsolidated basin-fill deposits and the poorly consolidated Salt Lake Formation underlying the basin

Specified-flux boundaries

- Infiltration of precipitation and unconsumed irrigation water
- Seepage from canals
- Seepage from streams
- Withdrawal from wells

Head-dependent flux boundaries

- Between the unconsolidated, saturated basin-fill deposits within the basin and the consolidated bedrock that bounds the basin
- Seepage to streams
- Spring discharge
- Evapotranspiration

The saturated, basin-fill deposits are active cells while the consolidated bedrock cells are inactive. Because the angle of the faults that bound the basin is very high, the same boundary between active and inactive cells is used in all layers, except in areas of shallow, consolidated rock. Where evidence of shallow consolidated rock is found, the lower layers are inactive. These areas of unconsolidated bedrock are found near Franklin, Idaho, and the confluence of Battle and Deer Creeks in Idaho.

Clarkston and Weston Canyons are not simulated by this model because both of these areas have their own individual basin-fill groundwater systems, and each is at a higher altitude than the main groundwater system in Cache Valley (Kariya et al. 1994).

Recharge

Robinson (1999) concluded that groundwater pumped from the principal aquifer is replaced by increased recharge. This is due to the water levels in the principal aquifer showing no long-term declines. Robinson (1999) explains this increased recharge as follows: in the unconfined portion of the principal aquifer, sufficient water infiltrates to maintain the head above the highest extent of the confining layers, and the groundwater flowing over the confining layers discharges to one of the many springs in Cache Valley. This indicates that the principal aquifer is in equilibrium. This is noted in the groundwater budget estimated for this study (Table 3).

Recharge to the groundwater system consists of infiltration of precipitation and unconsumed irrigation water, seepage from canals, seepage from streams, and other forms of recharge. Recharge to the groundwater system through infiltration of precipitation, unconsumed irrigation water, seepage from canals, and seepage from streams was simulated using the Recharge Package (McDonald and Harbaugh 1988).

Robinson (1999) describes the upper and lower confining layers as “clay grading to silt, sand, and gravel near the valley margins.” The recharge to the principal aquifer through infiltration of precipitation is most likely greater through the

Table 3: Estimated budget for the groundwater system in the unconsolidated basin-fill deposits in Cache Valley.

Recharge	**Flow (cubic feet per second)**
Net recharge of precipitation	140
Net recharge of unconsumed irrigation water	75
Seepage from canals	116
Seepage from streams	1
Total	332
Discharge	
Seepage to streams	55
Spring discharge	138
Evapotranspiration	87
Withdrawal from wells	52
Total	332

sand and gravel portions of the confining layers near the valley margin than through the clay and silt in the middle portion of the valley. The amount of recharge to the principal aquifer near the valley margins and in the middle portion of the valley is not known. It was therefore assumed that precipitation recharges the saturated, basin-fill deposits uniformly over the entire basin. The amount of recharge contributed through precipitation to the saturated, basin-fill deposits was divided evenly over the basin according to cell area.

An estimated 215 cubic feet per second (6.1 cubic meters per second) recharges the groundwater system through infiltration of precipitation and unconsumed irrigation water (Table 3). The average amount of precipitation that fell on Cache Valley from 1984-1997 is 1.2 feet per year (0.37 meters per year) (Utah State University 2002). This rainfall rate multiplied by the area of the study area (660 square miles) is equivalent to 700 cubic feet per second (20 cubic meters per second). Studies of other basins in Utah indicate that recharge to areas underlain by unconsolidated basin-fill deposits may range from 1 to 20 percent of the precipitation (Razem and Steiger 1981; Hood and Waddell 1968). It is assumed that 20 percent of the precipitation recharges the groundwater system. Therefore, 140 cubic feet per second (4 cubic meters per second) recharges the groundwater system through precipitation (Table 3 and Appendix). The service areas used in simulating recharge from unconsumed irrigation water were obtained from Kariya et al. (1994, Figure 19). Kariya et al. (1994) adapted the areas from maps made by the U.S. Soil Conservation Service (1976). Figure 6 displays the cells used in simulating recharge from unconsumed irrigation water. The amount of water that recharges the groundwater

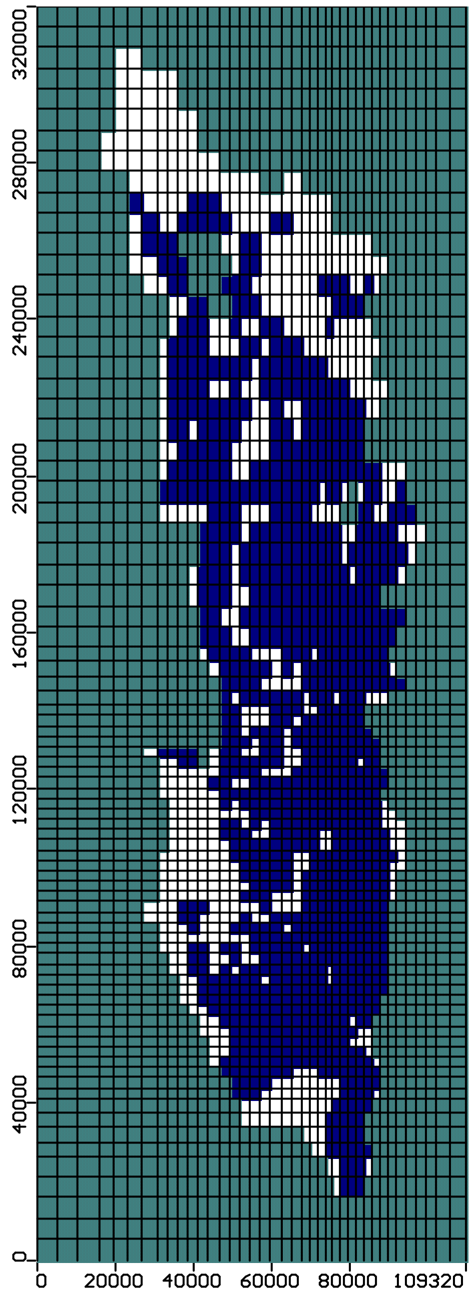

Figure 6: Location of cells representing recharge from unconsumed irrigation water (Kariya et al. 1994). White cells represent active cells, light shaded cells represent inactive cells, and dark shaded cells represent recharge from unconsumed irrigation water. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

system from unconsumed irrigation water was distributed evenly over the irrigated areas. As with precipitation, this was done according to cell area.

Recharge due to infiltration of unconsumed irrigation water was estimated using the following equation utilized by Kariya et al. (1994, p. 26).

$$R = [(TD \text{ X } CE)/IA] \text{ X } (1.0 - OE) \quad (1)$$

where

R is recharge for year of interest, in feet;
TD is total amount of water diverted to the area served by a canal, in acre-feet;
CE is canal conveyance efficiency estimated by the [U.S.] Soil Conservation Service (1976), in decimal form;
IA is irrigation company service area defined by the [U.S.] Soil Conservation service (1976), in acres;
OE is on-farm efficiency estimated by the [U.S.] Soil Conservation Service (1976), in decimal form.

The total amount of water diverted in canals is 283 cubic feet per second (10 cubic meters per second) (Kariya et al. 1994). Using an average canal conveyance efficiency of 59 percent, an average on-farm efficiency of 55 percent ([U.S.] Soil Conservation Service 1976), and an irrigated area of 158,835 acres, the recharge to the groundwater system is .34 feet per year (0.10 meters per year). This recharge rate multiplied by the area of 158,835 acres of farmland served by the canals results in 75 cubic feet per second (2.1 cubic meters per second) of recharge to the groundwater system (Table 3 and Appendix).

Recharge from seepage from canals was calculated using the following equation:

$$S = TD \text{ x } [(100\text{-}CE)/100] \quad (2)$$

where

S is canal seepage, in acre-feet;
TD is total amount of water diverted into a canal, in acre-feet;
CE is canal efficiency, expressed as a percent.

Equation 2, the total amount of water diverted into each canal, and the conveyance-efficiency estimates were all obtained from Kariya et al. (1994). All canals were assumed to have equivalent canal efficiencies. The amount of seepage was then divided by the total length of the canal to obtain the amount of seepage per foot of canal. This was then multiplied by the length of the canal within a certain cell to obtain the amount of seepage per cell. The locations of the major canals were obtained from Kariya et al. (1994, Plate 1) and are shown in Figure 7.

The amount of water that recharges the groundwater system through seepage from canals was calculated through multiplying the total amount of water diverted of 283 cubic feet per second (10 cubic meters per second) by 100 percent minus the average canal efficiency of 59 percent. This calculation estimated that 116 cubic feet per second (3.3 cubic meters per second) of water recharges the groundwater system through seepage from canals (Table 3).

Recharge to the groundwater system by seepage from streams was also simulated. High, Maple, and Mink Creeks are the creeks with the greatest losses, and were therefore simulated by the model. Kariya et al. (1994) contended that these streams are almost totally diverted near the mountain front during the irrigation season. Therefore, the first cells that the streams intercept as they flow through the basin-fill deposits were simulated by the model. The recharge from other streams in the valley are assumed to be small compared to the total recharge, and were not simulated by the model. The total amount of recharge to the groundwater system from

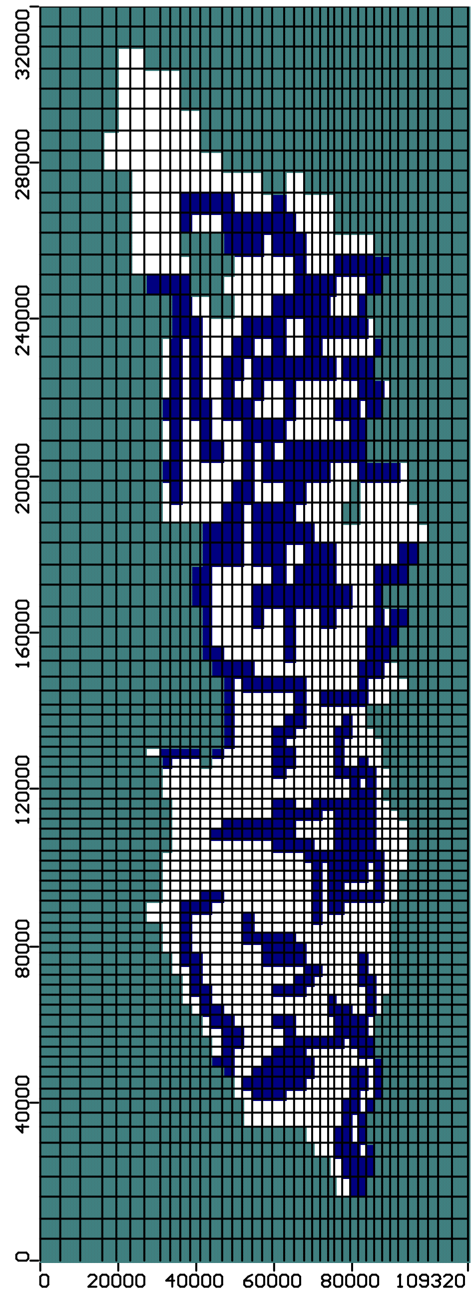

Figure 7: Location of cells representing recharge from canal seepage (Kariya et al. 1994). White cells represent active cells, light shaded cells represent inactive cells, and dark shaded cells represent recharge from canal seepage. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

seepage from streams is 1 cubic foot per second (0.03 cubic meters per second), which is equal to the difference between total discharge and recharge from infiltration of precipitation and unconsumed irrigation water and seepage from canals.

Discharge

Discharge from the groundwater system in the unconsolidated basin-fill deposits of Cache Valley includes seepage to streams, spring discharge, evapotranspiration, and withdrawal from wells.

Bjorklund and McGreevy (1971, p. 35) estimated the total discharge from both springs and seepage to streams to be 193 cubic feet per second (5.5 cubic meters per second). Kariya et al. (1994) performed a separate study on spring discharge. They concluded that 138 cubic feet per second (3.9 cubic meters per second) discharged from the groundwater system from springs (Table 2). Using Bjorklund and McGreevy's (1971) estimate of 193 cubic feet per second (5.5 cubic meters per second) for the combined discharge from springs and from seepage to streams, and subtracting the estimate by Kariya et al. (1994) of 138 cubic feet per second (3.9 cubic meters per second) for spring discharge alone yields a value of 55 cubic feet per second (1.6 cubic meters per second) for the discharge from the groundwater system as seepage to streams (Table 3).

Discharge from the groundwater system as seepage to streams was simulated using the River Package (McDonald and Harbaugh 1988). The River Package requires a river head conductance term as input for a river cell. The conductance is a numerical parameter that represents the resistance of flow across the river bed to the groundwater. The streambed conductance was calculated as:

$$C = [(KLW)/M] \tag{3}$$

where

K is the hydraulic conductivity of the river bed material, in feet per day;
L is the length of a reach through a cell, in feet;
W is the width of the river in the cell, in feet;
M is the thickness of the river bed, in feet.

The Little Bear, Logan, Blacksmith Fork, Cub, and Bear Rivers were simulated using the River Package. The hydraulic conductivity of the river bed material was assumed to be silty sand, which has an estimated value of 2.8 x 10^-2 feet per day (9.9 x 10^-6 centimeters per second) (Fetter 2001). For the Little Bear, Logan, Blacksmith Fork, and Cub Rivers, the thickness of the riverbed is assumed to be 4 feet (1.2 meters), and the width of each river to be 8 feet (2.4 meters). The thickness of the riverbed for the Bear River is assumed to be 10 feet (3 meters), and the width of the river to be 40 feet (12 meters). Figure 8 shows the location of river cells.

The model simulated spring discharge using the Drain Package (McDonald and Harbaugh 1988). A drain elevation and conductance are required as input for the Drain Package. The drain elevation was set at the elevation of the ground surface. There is no general formula for calculating drain conductance, due to the lack of detailed information required. This information includes detailed head distribution around the drain, aquifer hydraulic conductivity near the drain, distribution of fill material, number and size of the drain-pipe opening, the amount of clogging materials, and the hydraulic conductivity of clogging materials. It is common, with the proper selection of coefficients, to substitute the River Package conductance for the Drain

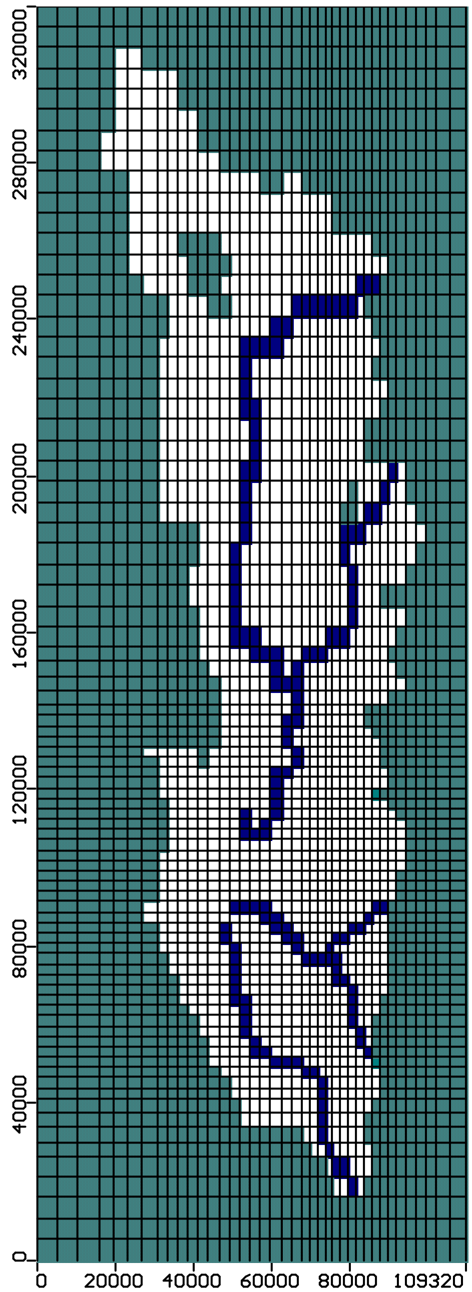

Figure 8: Location of cells that represent river boundary conditions (Kariya et al. 1994). White cells represent active cells, light shaded cells represent inactive cells, and dark shaded cells represent river boundary conditions. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

Package conductance (McDonald and Harbaugh 1988). This was the method used to calculate the drain conductance for this model. The hydraulic conductivity for the area around the drain was assumed to be the same as the value used for the river bed material, or 2.8 x 10^-2 feet per day (9.9 x 10^-6 centimeters per second). The width of the drains was set at 2 feet (0.6 meters), and the thickness of the drain beds was set at 20 feet (6 meters). The initial conductance of the cells representing springs ranged from 5.6 square feet per day (0.5 square meters per day) to 7.5 square feet per day (0.7 square meters per day). The locations of these cells were obtained from Kariya et al. (1994, Table 7), and are shown in Figure 9.

Kariya et al. (1994) estimated that 87 cubic feet per second (2.5 cubic meters per second) of water was lost from the groundwater system due to evapotranspiration. This value has been used for this model (Table 3).

Evapotranspiration was simulated using the Evapotranspiration Package (McDonald and Harbaugh 1988). The extinction depth for evapotranspiration cells was set at 6 feet (1.8 meters). The total amount of evapotranspiration from the basin was divided evenly over the cells where evapotranspiration was simulated. The locations of these cells were obtained from those areas designated by Bjorklund and McGreevy (1971, Plate 4), and are shown in Figure 10.

Through the study of well logs and pumping rates, Kariya et al. (1994) estimated that 52 cubic feet per second (1.5 cubic meters per second) was pumped from the groundwater system in 1990. This value has been used for this model (Table 3).

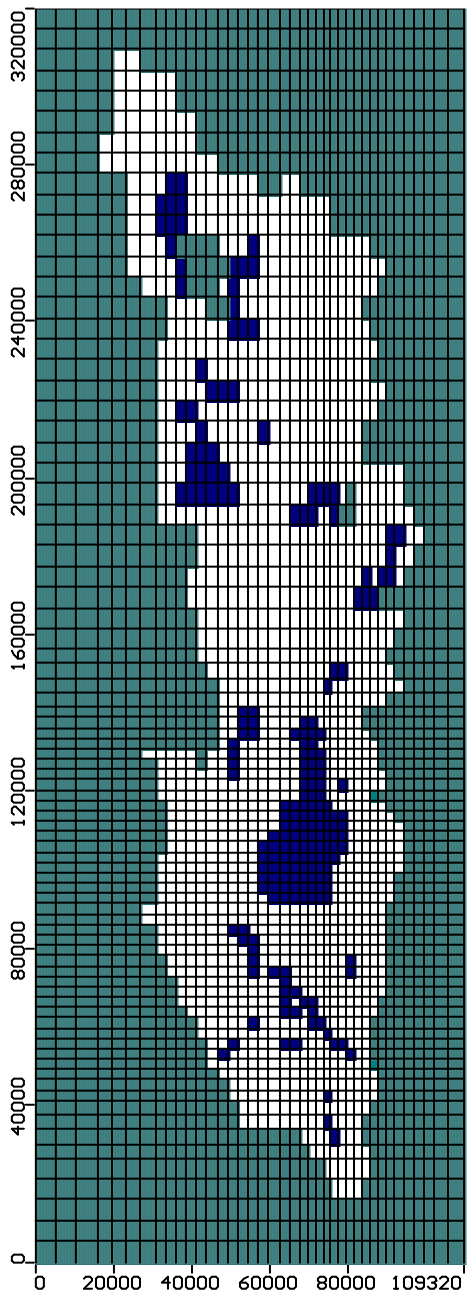

Figure 9: Location of cells representing springs (Kariya et al. 1994). White cells represent active cells, light shaded cells represent inactive cells, and dark shaded cells represent springs. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

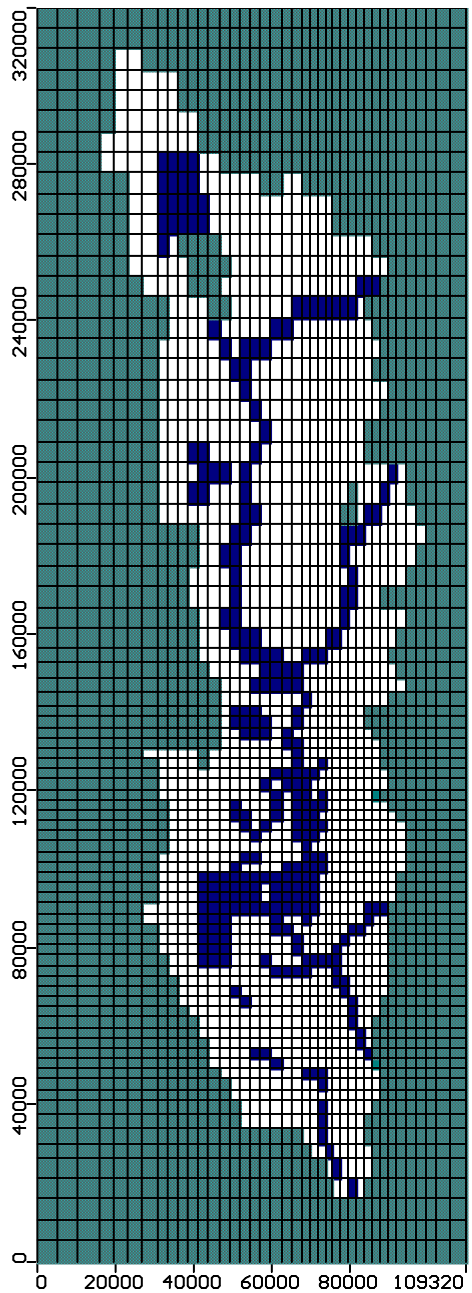

Figure 10: Location of cells representing discharge through evapotranspiration (Bjorklund and McGreevy 1971). White cells represent active cells, light gray cells represent inactive cells, and dark gray cells represent discharge through evapotranspiration. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

Discharge from the groundwater system from the pumping of wells was simulated using the Well Package (McDonald and Harbaugh 1988). One hundred sixteen pumping wells have been used in the model for simulation. Figure 11 shows that most of the pumping in Cache Valley occurs near the eastern margin of the valley within the principal aquifer. Kariya et al. (1994, Table 8) showed that 6 percent of the pumping is domestic, 46 percent of the pumping is municipal, and 48 percent of the pumping is for irrigation purposes. This ratio was used to dictate how much pumping was attributed to domestic, municipal, and irrigation purposes. The information regarding the location and amount of pumping from each well was obtained from the Utah Division of Water Rights (2002) and the Idaho Division of Water Resources (2002).

Hyrum and Cutler Reservoirs were simulated using the General Head Boundary Package (McDonald and Harbaugh 1988). The locations of the general head cells are shown in Figure 12. This package is mathematically similar to the River, Drain, and Evapotranspiration Packages (McDonald and Harbaugh 1988) in that flow into or out of a cell is head dependent. The head values for the general head cells representing Hyrum and Cutler reservoirs were initially set at 5 feet (1.5 meters) below the land surface. This value was chosen because the water level of the reservoirs is on average 5 feet (1.5 meters) below the land surface. The conductance of each cell was calculated using the conductance equation of the River Package (McDonald and Harbaugh 1988). The conductance of the corresponding cells range from 5.6 square feet per day (0.5 square meters per day) to 7.5 square feet per day (0.7 square meters per day). These calculated conductance values are the same as the

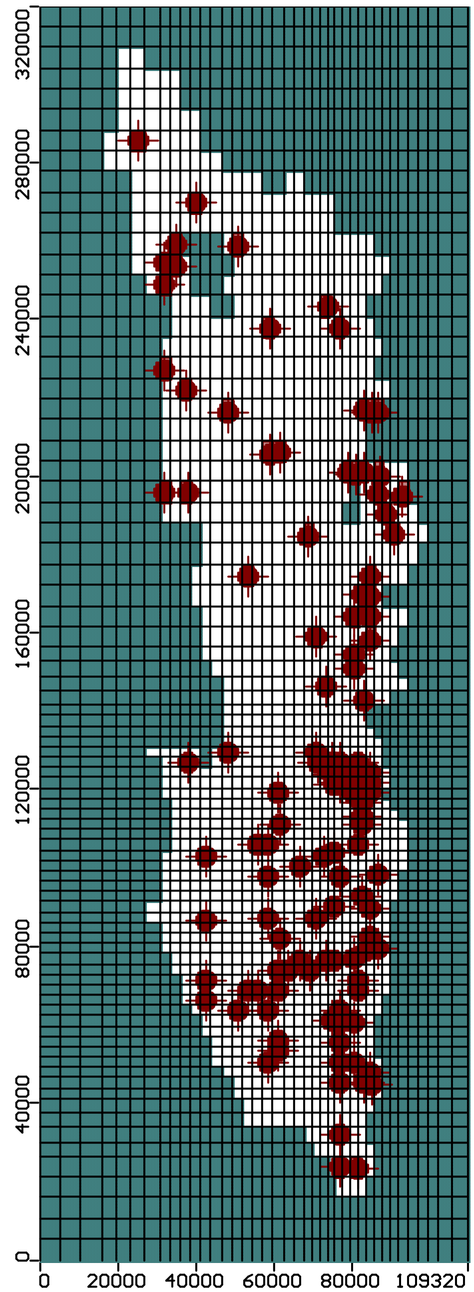

Figure 11: Location of cells representing discharge from well pumping. White cells represent active cells, light gray cells represent inactive cells, and dark gray circular cells represent discharge from well pumping. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

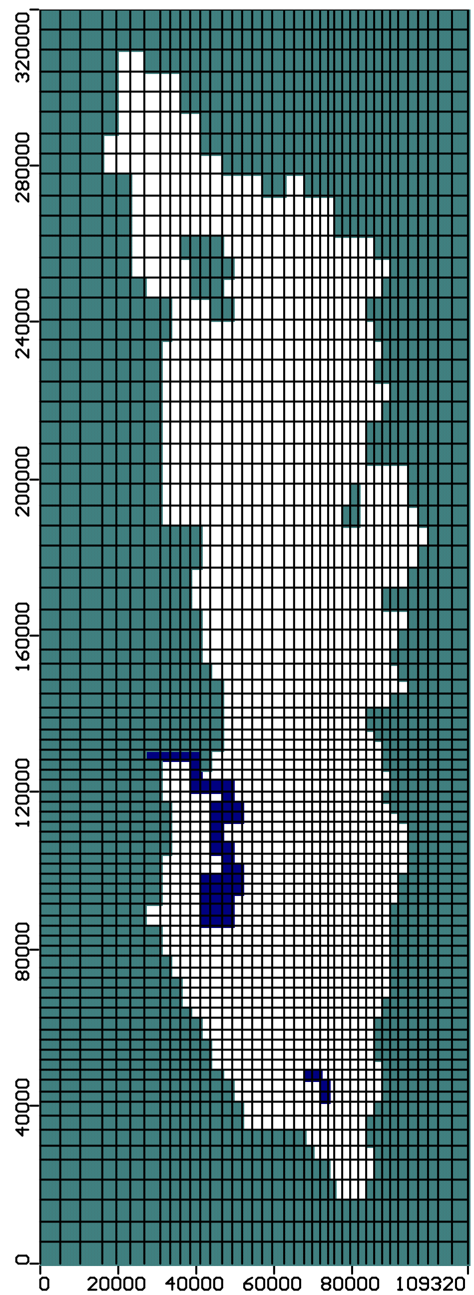

Figure 12: Location of cells representing Cutler and Hyrum Reservoirs. White cells represent active cells, light gray cells represent inactive cells, and dark gray cells represent the location of Hyrum and Cutler reservoirs. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

calculated conductance values for the spring cells. This is because the conductance equation of the River Package (McDonald and Harbaugh 1988) was used to calculate the conductance for each type of cell.

Hydraulic Properties

Several aquifer properties are required as input by MODFLOW (McDonald and Harbaugh 1988). These include the hydraulic conductivity, specific storage, specific yield, porosity, and the effective porosity. Hydraulic conductivity, porosity, effective porosity, and specific yield values were all obtained from Fetter (2001). The specific storage values were calculated using the equation given by Jacob (1940) and Cooper (1966). The compressibility of the aquifer values used in calculating specific storage values were obtained from Freeze and Cherry (1979).

Bjorklund and McGreevy (1971) divided Cache Valley into 11 different groundwater areas on the basis of their hydrologic parameters. The hydraulic properties were then assigned to the various areas according to the description of the aquifer material given by Bjorklund and McGreevy (1971), with four exceptions. Areas 3 and 4 were combined as one groundwater area, Areas 11 and 7 were also combined, and Areas 9 and 10 were combined as one groundwater area. This was done because the hydraulic properties necessary for input in MODFLOW (McDonald and Harbaugh 1988) did not vary significantly from one area to the other. Area 6 has not been simulated because it lies within the Clarkston bench, which was in an area of inactive cells because bedrock outcrops at the ground surface. The model therefore contains seven groundwater areas. These seven areas are shown in Figure 13.

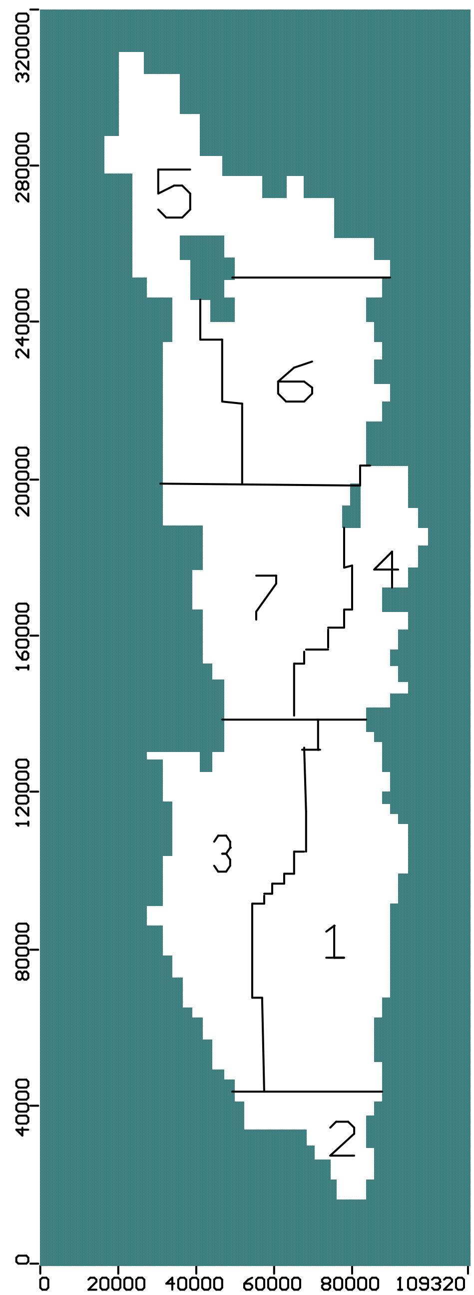

Figure 13: Location of the seven groundwater areas. Adapted from Bjorklund McGreevy (1971). The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

Horizontal and Vertical Hydraulic Conductivities

Robinson (1999, p. 27-28) describes the upper and lower confining layers as "clay grading to silt, sand, and gravel near the valley margins." The horizontal and vertical hydraulic conductivity values of the clay toward the center of the valley that have been used are 8 x 10^-4 feet per day (3 X 10^-7 centimeters per second) and 8 x 10^-5 feet per day (3 x 10^-8 centimeters per second), respectively, and the horizontal and vertical hydraulic conductivities of the silt and sand toward the valley margins used are 0.03 feet per day (1 X 10^-5 centimeters per second) and 0.003 feet per day (1 x 10^-6 centimeters per second), respectively, throughout all seven areas.

Robinson (1999, p. 26) also describes the upper confined aquifer as "gravels to cobbles interbedded with sand and silt. Clay beds present in discontinuous lenses." The upper confined aquifer consequently has been assigned horizontal and vertical hydraulic conductivity values of 14 feet per day (5 x 10^-3 centimeters per second) and 0.14 feet per day (5 x 10^-5 centimeters per second), respectively, throughout all seven areas.

Area 1 is described by Bjorklund and McGreevy (1971) as being the most productive aquifer in the valley. They describe the aquifer materials as being very coarse along the mountain front but becoming finer grained westward. Robinson (1999) describes the aquifer materials as thickly bedded gravels and sands with discontinuous lenses of silt and clay. A horizontal hydraulic conductivity of 100 feet per day (3.5 x 10-^2 centimeters per second) and a vertical hydraulic conductivity of 1 x 10^-3 feet per day (3.5 x 10^-7 centimeters per second) were assigned to both the unconfined aquifer and the lower confined aquifer.

Bjorklund and McGreevy (1971) describe Area 2 as having thin deposits of gravel overlying mostly fine-grained material. The unconfined aquifer was consequently assigned a horizontal hydraulic conductivity of 40 feet per day (1.4 x 10^-2 centimeters per second) and a vertical hydraulic conductivity of 38 feet per day (1.3 x 10^-2 centimeters per second). The lower confined aquifer has been assigned a horizontal hydraulic conductivity of 3 feet per day (1 x 10 ^-3 centimeters per second) and a vertical hydraulic conductivity of 3 x 10^ -4 feet per day (1 x 10^-7 centimeters per second).

Area 3 is composed predominantly of clay and silt with thin beds of sand and fine gravel (Bjorklund and McGreevy 1971). This applies to both the unconfined and the lower confined aquifers. Accordingly, a horizontal conductivity of 3 feet per day (1 x 10^-3 centimeters per second) and a vertical hydraulic conductivity of 3 x 10^-4 feet per day (1 x 10^-7 centimeters per second) were applied to both the unconfined and lower confined aquifer.

Interbedded clay, silt, sand, and gravel overlie the Tertiary conglomerate in most of Area 4 (Bjorklund and McGreevy 1971). The unconfined and lower confined aquifers consequently have been assigned a horizontal hydraulic conductivity of 15 feet per day (5 x 10^-3 centimeters per second) and a vertical hydraulic conductivity of 0.01 feet per day (3 x 10^-6 centimeters per second).

Permeable deposits of sand and gravel containing both confined and unconfined groundwater are mainly what constitute Area 5 (Bjorklund and McGreevy 1971). A horizontal hydraulic conductivity of 10 feet per day (3 x 10^-3 centimeters per second) and a vertical hydraulic conductivity of 0.08 feet per day (3 x 10^-5 centimeters per second) were assigned to the unconfined and lower confined aquifers of Area 5.

Area 6 is found on the high bench area north and south of the Bear River near Preston. It is composed of thick deposits of fine sand and silt (Bjorklund and McGreevy 1971). The horizontal and vertical hydraulic conductivities used in this area are 1.4 feet per day (4.9 x 10^-4 centimeters per second) and 0.05 feet per day (2 x 10^-5 centimeters per second), respectively. These values were applied to the unconfined and lower confined aquifers.

According to Bjorklund and McGreevy (1971), Area 7 is composed of sand and silt, which cover lake-bottom clays. The horizontal and vertical hydraulic conductivities of the unconfined aquifer used are therefore 1 foot per day (3.5 x 10^-4 centimeters per second) and 0.05 feet per day (2 x 10^-5 centimeters per second), respectively, while the horizontal and vertical hydraulic conductivities of the lower confined aquifer used are 3 x 10^-4 feet per day (1 x 10^-7 centimeters per second) and 3 x 10^-5 feet per day (1 x 10^-8 centimeters per second), respectively.

Specific Storage, Specific Yield, Porosity, and Effective Porosity Values

The values for specific storage, specific yield, porosity, and effective porosity for confined and unconfined conditions are shown in Table 4 for the seven groundwater areas. The values for specific storage, specific yield, porosity, and effective porosity for the continuous confining layers and the upper confined aquifer are shown in Table 5.

Fetter (2001) suggests that, at least in sediments, all the pores are connected, and there is no need to be concerned with the effective porosity with respect to flow of water. However, MODFLOW (McDonald and Harbaugh 1988) requires a value for effective porosity. Consequently, the values for porosity and effective porosity were considered to be equal for each individual area.

Table 4: Values for the specific storage, specific yield, porosity, and effective porosity for the lower confined and unconfined aquifers for the seven groundwater areas in the unconsolidated basin-fill deposits in Cache Valley.

		Confined Aquifer	Unconfined Aquifer
Area 1	Specific Storage (ft^-1)	2 x 10^-6	2.5 x 10^-3
	Specific Yield	0.25	0.25
	Porosity	0.27	0.27
	Effective Porosity	0.27	0.27
Area 2	Specific Storage (ft^-1)	2 x 10^-5	2.5 x 10^-3
	Specific Yield	0.23	0.23
	Porosity	0.33	0.44
	Effective Porosity	0.33	0.44
Area 3	Specific Storage (ft^-1)	2 x 10^-5	1.3 x 10^-3
	Specific Yield	0.13	0.13
	Porosity	0.44	0.44
	Effective Porosity	0.44	0.44
Area 4	Specific Storage (ft^-1)	3 x 10^-5	1.9 x 10^-3
	Specific Yield	0.19	0.19
	Porosity	0.41	0.41
	Effective Porosity	0.41	0.41
Area 5	Specific Storage (ft^-1)	2 x 10^-6	2.4 x 10^-3
	Specific Yield	0.24	0.24
	Porosity	0.35	0.35
	Effective Porosity	0.35	0.35
Area 6	Specific Storage (ft^-1)	4 x 10^-5	2 x 10^-3
	Specific Yield	0.2	0.2
	Porosity	0.4	0.4
	Effective Porosity	0.4	0.4
Area 7	Specific Storage (ft^-1)	4 x 10^-5	2.2 x 10^-3
	Specific Yield	0.22	0.22
	Porosity	0.4	0.4
	Effective Porosity	0.4	0.4

Table 5: Values for the specific storage, specific yield, porosity, and effective porosity for the continuous confining layers and the upper confined aquifer in the unconsolidated basin- fill deposits in Cache Valley.

		Clay	Silt	Sand
Confining Layers	Specific Storage (ft^-1)	2 x 10^-5	6 x 10^-5	1 x 10^-6
	Specific Yield	0.07	0.18	0.26
	Porosity	0.46	0.42	0.26
	Effective Porosity	0.46	0.42	0.26
Upper Confined Aquifer	Specific Storage (ft^-1)	2 x 10^-5		
	Specific Yield	0.22		
	Porosity	0.38		
	Effective Porosity	0.38		

CHAPTER IV

MODEL CALIBRATION AND SIMULATIONS

Steady-State Calibration

Robinson (1999) found that the water levels in the principal aquifer show no long-term declines. Robinson (1999) explains that in the unconfined portion of the principal aquifer, there is sufficient recharge to maintain the head above the top of the easternmost extent of the upper confining layers. Consequently, the groundwater flowing over the top of the upper confining layer becomes part of the unconfined aquifer, and discharges in one of the many springs in Cache Valley. This indicates that the principal aquifer is in equilibrium. The model created for this thesis was therefore calibrated to 1999 conditions, which have been assumed to be steady-state.

The model was considered to have reached steady-state when the simulated drawdown within the various aquifers remained constant. Figure 14 depicts the drawdown in various locations throughout the valley in both the unconfined and lower confined aquifers over a time period of 30 years. Figure 14 shows that the model reached steady-state after approximately three years.

The head data for the unconsolidated basin-fill deposits were obtained from Robinson (1999) and the Idaho Division of Water Resources (2002). It is important to note that, due to the lack of consistent well information, the head values used in calibrating the model were not all taken at the same time of the year or even within the same year.

Fifty-five observation wells were used in calibrating the model (Figure 15).

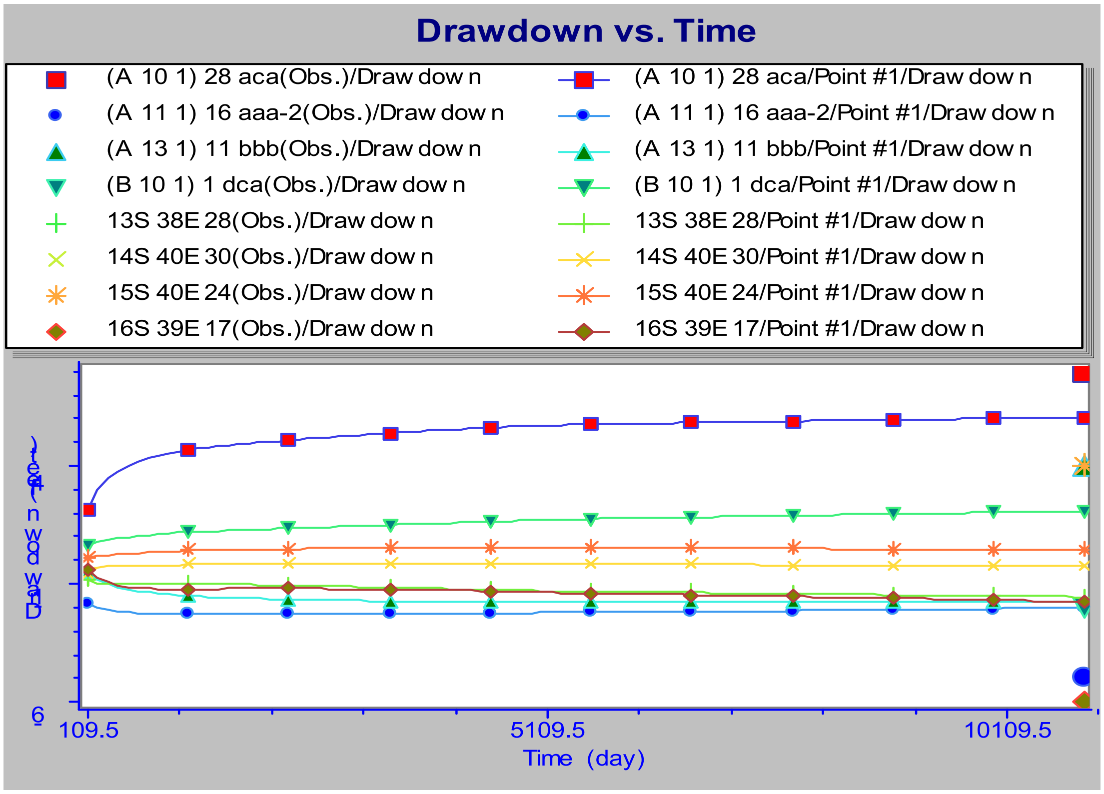

Figure 14: Drawdown versus time for selected head observation wells located throughout the valley in both the unconfined and lower confined aquifers over a time period of 30 years.

Of these wells, 27 are located in the unconfined aquifer, 15 are located in the upper confined aquifer, and 13 are located in the lower confined aquifer. All 55 of the observation wells are independent of the 116 pumping wells. The model was calibrated such that the observed heads of the observation wells are within 6 feet (1.8 meters) of the heads calculated by the model. Figure 16 is a graph showing the relationship between calculated and observed heads. There were no trends in correlation to the screened interval of the observation wells.

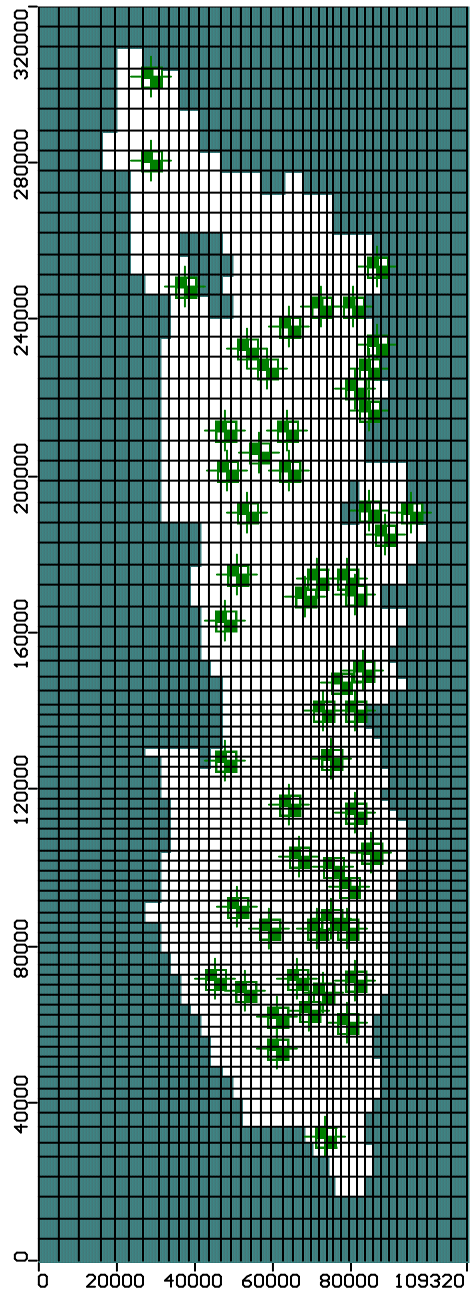

Figure 15: Location of observation wells used to calibrate heads. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

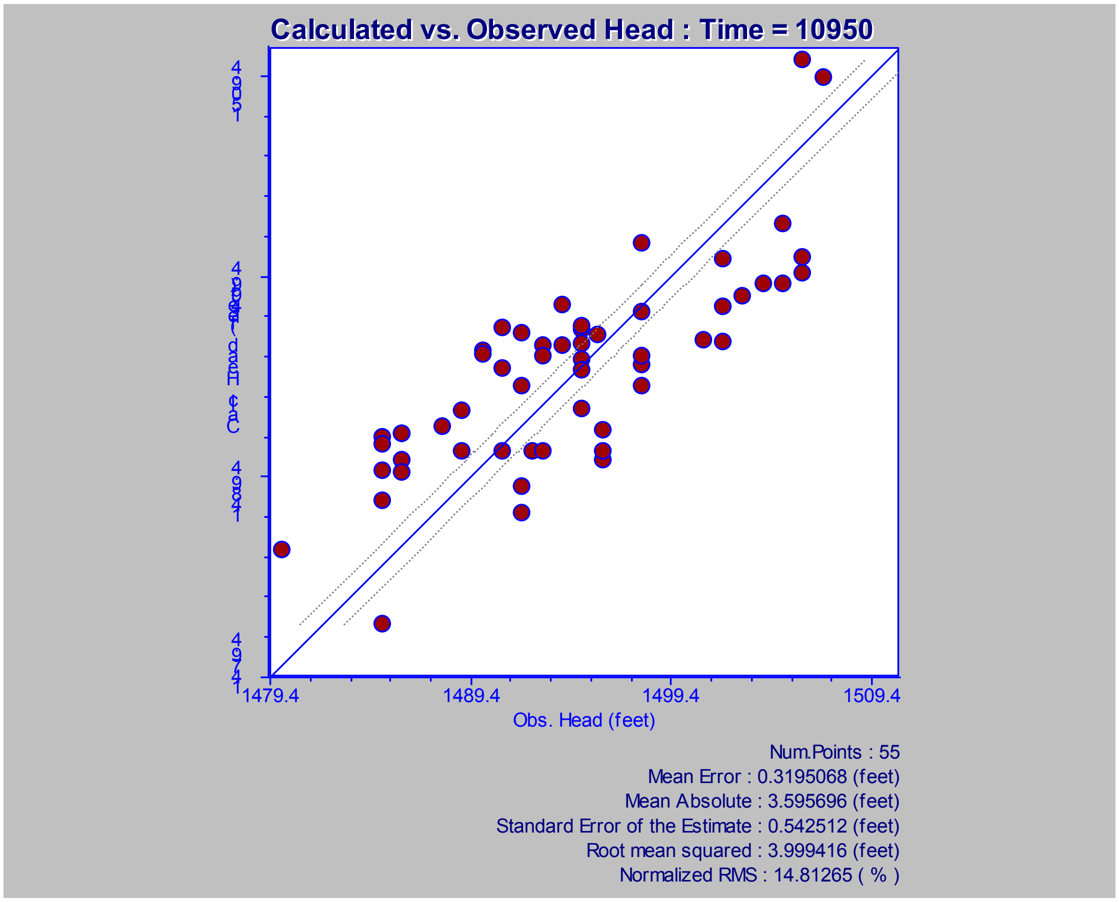

Figure 16: Graph showing relationship between calculated and observed heads.

Many springs are located in cells that are adjacent to river cells. In order to simulate discharge from the groundwater system through springs, it was necessary to produce heads that were at or above the ground surface. With the head at or above the ground surface, very little water was discharging to the streams. It therefore became necessary to simulate seepage to streams using the Drain Package (McDonald and

Harbaugh 1988). In this manner, the model was capable of discharging sufficient water from the groundwater system to the streams.

During calibration, heads that were 10-15 feet (3.1-4.5 meters) above the ground surface were observed in various areas throughout the valley. In order to reduce these heads, the horizontal hydraulic conductivity of the cells in the area was set at 100 feet per day (3.5 x 10^-2 centimeters per second) to dissipate the water laterally.

General head boundary cells were also introduced along the margins of the model in the unconfined aquifer to assist in lowering the heads (Figure 17). Upon final steady-state calibration, conductance values of the general head boundary cells were 5,000 square feet per day (464.5 square meters per day). In order to dissipate the amount of water necessary to achieve the desired heads, 69 cubic feet per second (2.0 cubic meters per second) of water was discharged through the general head boundary cells. In order to maintain an equivalent amount of total discharge from the groundwater system, discharge from the groundwater system through springs was reduced by 70 cubic feet per second (2.0 cubic meters per second).

During calibration, the model-computed heads in the lower confined aquifer in the Utah portion of the valley were lower than the observed heads. Various parameters, such as the vertical hydraulic conductivity of the two continuous confining layers, the horizontal and vertical conductivities of the lower confined aquifer, and the specific storage and porosity of the lower confined aquifer, were all altered to increase the model-computed heads in the lower confined aquifer. Altering these parameters did not sufficiently increase the heads in the lower conifed aquifer.

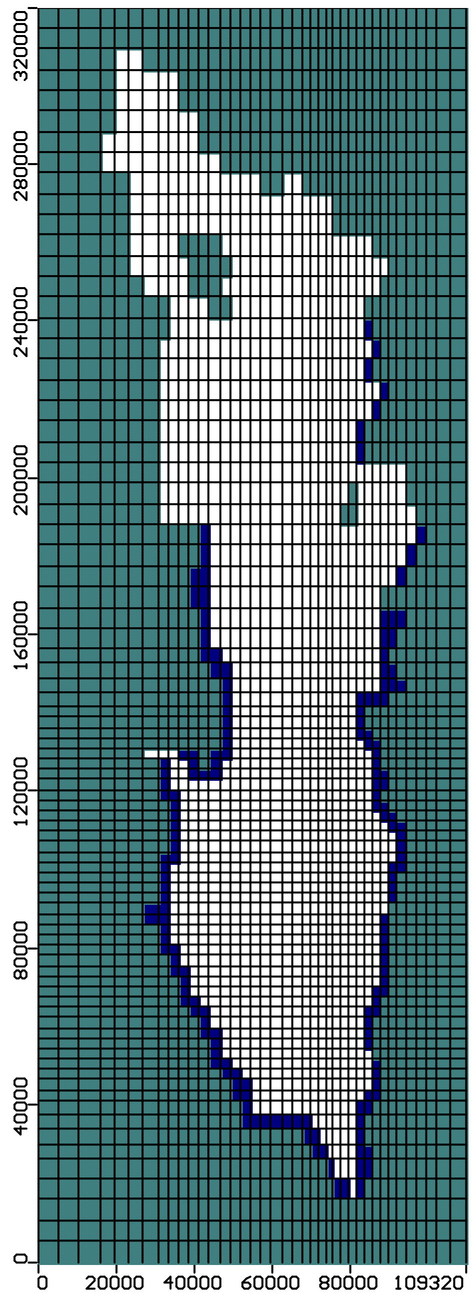

Figure 17: Location of general head boundary cells in the unconfined aquifer. White cells represent active cells, light gray cells represent inactive cells, and dark gray cells represent the location of the general head boundary cells. The units on each of the axes represent the length and width of the unconsolidated basin-fill deposits of Cache Valley in feet.

Therefore, it was also necessary to simulate 63 cubic feet per second (1.8 cubic meters per second) of recharge into the lower confined aquifer from the eastern and western margins of the valley to obtain the correct model-computed heads (Figure 18). Recharge to the lower confined aquifer from the western margin of the valley was simulated in Layers 2 through 4. Recharge to the lower confined aquifer from the eastern margin of the valley was simulated in Layers 2 through 8. Upon final steady-state calibration, the conductance of the general head boundary cells simulating recharge from the western margin of the valley was 500 square feet per second (46.5 square meters per second), and the conductance of the general head boundary cells simulating recharge from the eastern margin of the valley was 1,000 square feet per second (92.9 square meters per second). In order to maintain an equivalent amount of total recharge to the groundwater system, the total recharge to the groundwater system due to infiltration of precipitation and unconsumed irrigation water, seepage from canals, and seepage from streams was reduced by 63 cubic feet per second (1.8 cubic meters per second).

The conductance and drain elevations of spring and river cells and the evaportranspiration rate and evapotranspiration elevations of the cells representing evapotranspiration were all changed during calibration in an attempt to discharge the correct amount of water from the groundwater system through springs, rivers, and evapotranspiration. Altering these parameters of spring, river, and evapotranspiration cells could not discharge the necessary amount of water from the groundwater system. It therefore became necessary to increase the amount of recharge in cells representing springs, rivers, and evapotranspiration to achieve the correct amount of discharge

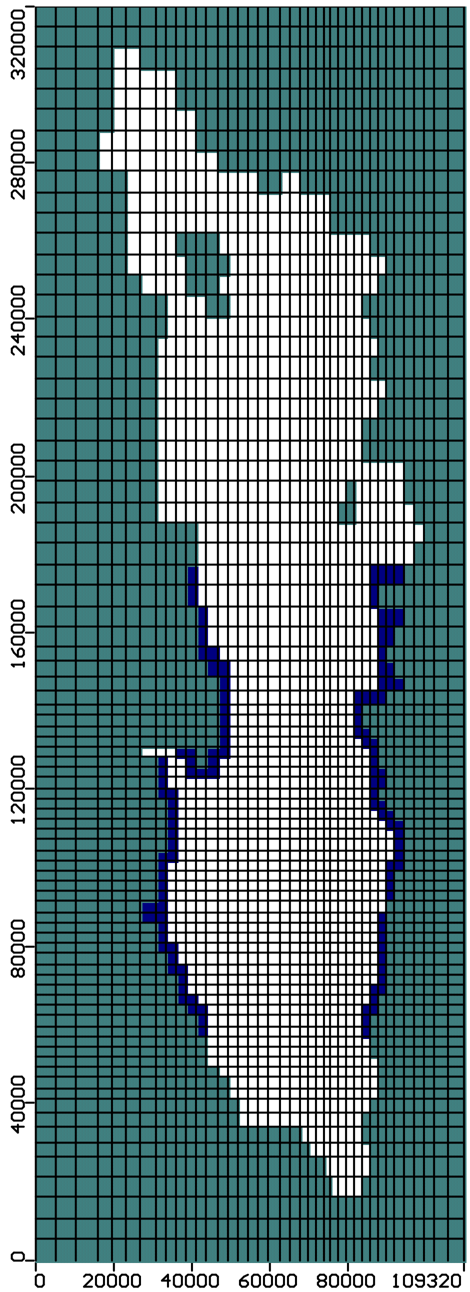

Figure 18: Location of general head boundary cells in layers two through four Along the western margin of the valley and in layers two through eight along the eastern margin of the valley. White cells represent active cells, light gray cells represent inactive cells, and dark gray cells represent the location of the general head boundary cells. The units on each of the axes represent the length and width of the unconsolidated basin-fill depostis of Cache Valley in feet.

through springs, rivers, and evapotranspiration. Consequently, an equivalent amount of recharge was reduced in cells that did not represent discharge from the groundwater system.

One aspect of the steady-state calibration was to compare model-computed fluxes with measured and estimated fluxes. Model-computed and estimated groundwater budgets are presented in Table 6. Estimated recharge from infiltration of precipitation and unconsumed irrigation water, seepage from canals, and seepage from streams is similar to simulated recharge for these components because they were simulated using specified-flux boundary conditions. The model-computed flux for seepage from streams was equal to 0.04 cubic feet per second (0.001 cubic meters per second), but this value is so small that it was considered to be negligible (Table 6).

Estimated discharge from withdrawal from wells is also similar to simulated discharge for these components because it is simulated using specified-flux boundary conditions. Model-computed discharge from seepage to streams, evapotranspiration, and spring discharge was calibrated to be nearly equal to the estimated discharge for each of the respective constituents.

Sensitivity Analysis

A sensitivity analysis was performed to determine the sensitivity of the model to the values used for the various hydraulic parameters. A sensitivity analysis also provides insight into the magnitude of error that may be associated with values of

Table 6: Estimated and model-computed groundwater budgets for the steady-state simulation.

Recharge	**Estimated Flow (cubic feet per second)**	**Model-Computed Flow (cubic feet per second)**
Net recharge of precipitation	140	108
Net recharge of unconsumed irrigation water	75	75
Seepage from canals	116	85
Seepage from streams	1	0
General Head Boundaries	0	63
Total	332	331
Discharge		
Seepage to streams	55	54
Spring discharge	138	68
Evapotranspiration	87	85
Withdrawal from wells	52	53
General Head Boundaries	0	69
Total	332	329

poorly known hydraulic parameters. The sensitivity analysis was performed during the steady-state calibration of the model.

Of the various hydraulic properties, the vertical hydraulic conductivity of the confining layers proved to produce the greatest amount of head change. During calibration, the vertical hydraulic conductivity of the confining layers was altered to assist in calibrating heads in the lower confined aquifer. Changes in head remained constant whether the vertical hydraulic conductivity of the confining layers was altered in the center of the valley or towards the valley margins. Changing the vertical hydraulic conductivity value, within the range of values specified by Freeze and Cherry (1979) for the various types of material (2.8 x 10^-6 feet per day to 2.8x 10^-3 feet per day), would change heads in the lower confined aquifer by a magnitude of 1-2 feet (0.3-0.6 meters). Through first simulating recharge to the lower confined aquifer from the Wellsville Mountains and the Bear River Range, the model computed heads were found to be comparable to the observed heads. The vertical hydraulic conductivity of the confining layers was then altered to calibrate the model-computed heads to within 6 feet (1.8 meters) of the observed heads.

Altering the vertical hydraulic conductivity values of the various aquifers, within the range of values specified by Fetter (2001) for the various types of aquifer material (2.8x 10^-3 feet per day to 2.8 feet per day), would produce average head changes of 0-1 feet (0-0.3 meters). As with the confining layers, changes in head remained constant whether the vertical hydraulic conductivity of the various aquifers was altered in the center of the valley or towards the valley margins. The remaining hydraulic properties, namely the specific storage, porosity, effective porosity, and the specific yield, did not produce any significant head changes, and were altered very little during calibration.

The model-computed water levels in the unconfined aquifer were more sensitive to the head of the cells representing springs, rivers, and evapotranspiration than to the conductance or evapotranspiration rate of the respective cells. A large modification of the conductance value or evapotranspiration rate was required in order to produce the desired changes in discharge from the groundwater system, while a change of 1 foot (0.3 meters) in the head of the respective cells would produce a change in discharge from the groundwater system of up to 6 cubic feet per second (0.2 cubic meters per second). In cells where springs, rivers, or evapotranspiration were represented together, discharge from the groundwater system from the various constituents was mainly regulated by altering the head of the cells.

First Transient Simulation

Two transient-state simulations were run. The first simulation was run for 30 years with average precipitation and increased pumping from the principal aquifer. The Utah State Engineer has limited future increases in pumpage to 35 cubic feet per second (1 cubic meter per second). Consequently, 20 pumping cells were evenly spaced along the eastern margin of the principal aquifer, and each has been pumped at a rate of 1.8 cubic feet per second (0.05 cubic meters per second). The majority of the wells in the principal aquifer are completed in Layers 5 and 6. Consequently, the increased pumping was simulated in Layers 5 and 6 of the model.

Transient-state model-computed fluxes were compared to steady-state fluxes. Transient-state model-computed and steady-state groundwater budgets are presented in Table 7. Recharge through general head boundaries from the Wellsville Mountains and the Bear River Range increased by 17 cubic feet per second (0.5 cubic meters per second) (Table 7). Discharge from withdrawal from wells is 34 cubic feet per second (1 cubic meter per second) greater than the steady-state discharge. This is due to the addition of the 20 additional pumping cells in the principal aquifer.

Increased pumping in the principal aquifer had little effect on the unconfined aquifer of Cache Valley. Model-computed discharge from the groundwater system through springs, seepage to streams, evapotranspiration, and general head boundaries remained unchanged when compared with the model computed water budget for the steady-state simulation (Table 7). This suggests that the continuous confining layers that separate the unconfined aquifer and the principal aquifer may serve as an effective barrier between the two aquifers, and impede the flow of groundwater between them.

The increase of 34 cubic feet per second (1 cubic meters per second) in groundwater withdrawal through increased pumping caused an increase of 17 cubic feet per second (0.5 cubic meters per second) of recharge to the groundwater system through general head boundary cells. The general head boundary cells located along the eastern margin of the valley in Layers 2 through 8 are the cells in which the increased recharge occurred. The remaining 17 cubic feet per second (0.5 cubic meters per second) is accounted for by the discrepancy between the total recharge (+17 cubic feet per second) and the total discharge (+34 cubic feet per second).

Table 7: Steady-state and transient-state model-computed budgets for the transient-state simulation with increased pumping and average recharge.

Recharge	**Steady-state model-computed flow (cubic feet per second)**	**Transient-state model-computed flow with average recharge (cubic feet per second)**
Net recharge of precipitation	108	108
Net recharge of unconsumed irrigation water	75	75
Seepage from canals	85	85
Seepage from streams	0	0
General Head Boundaries	63	80
Total	331	348
Discharge		
Seepage to streams	54	54
Spring discharge	68	68
Evapotranspiration	85	85
Withdrawal from wells	53	87
General Head Boundaries	69	69
Total	329	363

This 17 cubic feet per second (0.5 cubic meters per second) discrepancy is associated with the general head boundary cells along the western margin of the valley. The increased pumping within the principal aquifer did not stimulate increased recharge along the western margin of the valley. This shows that the model is not reacting to the increased pumping in the same way that the natural system would react. This discrepancy provides less confidence in the transient-state simulation than with the steady-state simulation. The transient simulation reached steady-state in approximately three years, which corresponds to the amount of simulated time for the steady-state simulation to reach steady-state. True steady-state conditions would take longer to achieve due to the amount of time required for the increased pumping cells to stimulate increased recharge from the western margin of the valley. Therefore, true steady-state conditions for the transient simulation were not truly reached in three years.

Second Transient Simulation

The second simulation was also run for 30 years with the same increased pumping rate distributed evenly among the same 20 wells as the first simulation. This simulation differs in that it simulates less than average precipitation. One foot per year (0.3 meters per year), the lowest annual precipitation rate that fell on Cache Valley from 1984-1997, was used in the second simulation (Utah State University 2002).

Transient-state model-computed fluxes with decreased recharge were compared with steady-state fluxes. Transient-state model-computed and steady-state groundwater budgets are presented in Table 8. Estimated recharge from infiltration of unconsumed irrigation water, seepage from canals, and seepage from streams is similar to simulated recharge for these components because they were simulated with specified-flux boundary conditions.

Estimated recharge from infiltration of precipitation is 22 cubic feet per second (0.62 cubic meters per second) less than the steady-state recharge. This is a result of the decrease in the amount of precipitation in order to simulate less than average precipitation. Estimated recharge from the general head boundaries increased by 17 cubic feet per second (0.48 cubic meters per second), though, resulting in a net decrease in total recharge of only five cubic feet per second (0.14 cubic meters per second).

Estimated discharge from withdrawal from wells is 34 cubic feet per second (1 cubic meter per second) more than the steady-state discharge. This is due to the installation of the 20 additional pumping wells in the principal aquifer.

Decreasing recharge to the groundwater system through infiltration of precipitation affected discharge from the groundwater system through springs, seepage to streams, evapotranspiration, and general head boundaries. Discharge from the groundwater system through seepage to streams, springs, evapotranspiration, and general head boundaries decreased by three cubic feet per second (0.08 cubic meters per second), six cubic feet per second (0.17 cubic meters per second), five cubic feet per second (0.14 cubic meters per second), and three cubic feet per second (0.08 cubic meters per second), respectively.

Table 8: Steady-state and transient-state model-computed budgets for the transient-state simulation with increased pumping and less than average recharge.

Recharge	**Steady-state model-computed flow (cubic feet per second)**	**Transient-state model-computed flow with less than average recharge (cubic feet per second)**
Net recharge of precipitation	108	86
Net recharge of unconsumed irrigation water	75	75
Seepage from canals	85	85
Seepage from streams	0	0
General Head Boundaries	63	80
Total	331	326
Discharge		
Seepage to streams	54	51
Spring discharge	68	62
Evapotranspiration	85	80
Withdrawal from wells	53	87
General Head Boundaries	69	66
Total	329	346

A discrepancy of 20 cubic feet per second (0.57 cubic meters per second) exists between the total recharge and total discharge. Five cubic feet per second (0.14 cubic meters per second) of discrepancy is associated with the net decrease in total recharge, while 17 cubic feet per second (0.48 cubic meters per second) of discrepancy is associated with the net increase in total discharge. The remaining two cubic feet per second (0.06 cubic meters per second) discrepancy is associated with the discrepancy between the total recharge and total discharge of the steady-state model-computed flows. The increased pumping within the principal aquifer did not stimulate increased recharge along the western margin of the valley. This shows that the model is not reacting to the increased pumping in the same way that the natural system would react. This discrepancy provides less confidence in the second transient-state simulation than with the steady-state simulation. The second transient simulation reached steady-state in approximately three years, which corresponds to the amount of simulated time for the steady-state simulation to reach steady-state. True steady-state conditions would take longer to achieve due to the amount of time required for the increased pumping cells to stimulate increased recharge from the western margin of the valley. Therefore, true steady-state conditions for the second transient simulation were not truly reached in three years.

CHAPTER V

CONCLUSIONS, LIMITATIONS, AND SUGGESTIONS FOR FURTHER WORK

Conclusions

Calibration

During steady-state calibration, the model showed that recharge to the lower confined aquifer may occur along the eastern and western margins of Cache Valley. The simulated inflow along the eastern margin of the valley extends south from Richmond to Hyrum. The simulated inflow along the western margin of the valley extended south from Cornish to Wellsville. The simulation of subsurface inflow from the Wellsville Mountains and the Bear River Range was necessary to produce model-computed heads that were comparable to observed heads. This suggests that the natural system receives recharge to the principal aquifer from the surrounding mountain ranges.

Calibration also demonstrated that discharge from the unconfined aquifer may occur along the eastern and western margins of the valley. The simulated outflow along the eastern margin of the valley extended south from Riverdale to just south of Preston, and south from Lewiston to the southern end of the valley. The simulated outflow along the western margin of the valley extended south from Cornish to the southern end of the valley. The simulation of subsurface outflow from the unconfined aquifer was necessary to produce model-computed heads that were comparable to observed heads. This suggests that the natural system discharges groundwater from the unconfined aquifer near the valley margins.

Fifty-five observation wells were used in the steady-state calibration. The model was calibrated such that the greatest difference between the observed heads of the observation wells and the heads calculated by the model was 6 feet (1.8 meters).

Another aspect of the steady-state calibration was to compare model-computed fluxes with measured and estimated fluxes. Model-computed total recharge to the groundwater system was 1 cubic foot per second (0.03 cubic meters per second) less than the estimated total recharge to the groundwater system, and the model-computed total discharge from the groundwater system was 3 cubic feet per second (0.08 cubic meters per second) less than the estimated total discharge from the groundwater system.

During steady-state calibration, model computed fluxes of total recharge to the groundwater system and total discharge from the groundwater system were compared to establish when the model had reached steady-state. Steady-state conditions were achieved after approximately three years. The model computed total recharge was 2 cubic feet per second (0.06 cubic meters per second) greater than the model-computed total discharge. This discrepancy represents less than a 1% difference between total recharge to the groundwater system and total discharge from the groundwater system.

Predictive Simulations

Two simulations were run with increased pumping of 34 cubic feet per second (1 cubic meter per second) from the principal aquifer. Twenty pumping cells along the eastern margin of the major aquifer were used to simulate the increased pumping. The increased pumping was simulated in Layers 5 and 6 of the model. The first simulation was run with the average annual precipitation value of 1.2 feet per year

(0.36 meters per year), while the second was run with a value of 1 foot per year (0.3 meters per year) to simulate the lowest measured during the period from 1984 to 1997.

The first simulation produced very little change within the unconfined aquifer. The discharge from the groundwater system through springs, seepage to streams, evapotranspiration, and general head boundaries remained unchanged with the increase in discharge through pumping. This indicates that the two continuous, confining layers that blanket the valley may serve as a barrier to groundwater flow between the unconfined and lower confined aquifer. The increased pumping caused an increase of 17 cubic feet per second (0.5 cubic meters per second) of recharge in the general head boundary cells that simulate recharge to the groundwater system through subsurface inflow from the Bear River Range. However, the remaining 17 cubic feet per second (0.5 cubic meters per second) of increased discharge due to pumping represents the discrepancy between total recharge and total discharge.

This discrepancy of 17 cubic feet per second (0.5 cubic meters per second) is associated with the general head boundary cells along the western margin of the valley. The increased pumping within the principal aquifer did not stimulate increased recharge along the western margin of the valley. This shows that the model is not reacting to the increased pumping in the same way that the natural system would react. True steady-state conditions would take longer than three years to achieve due to the amount of time required for the increased pumping cells to stimulate increased recharge from the western margin of the valley.

During the second simulation, decreased recharge to the groundwater system through infiltration of precipitation affected discharge from the groundwater system through seepage to streams, springs, evapotranspiration, and general head boundaries. Model-computed discharge from the groundwater system through springs, seepage to streams, evapotranspiration, and general head boundaries decreased.

A discrepancy of 20 cubic feet per second (0.57 cubic meters per second) exists between the total recharge and total discharge. The majority of the discrepancy is associated with the general head boundary cells along the western margin of the valley. The increased pumping within the principal aquifer did not stimulate increased recharge along the western margin of the valley. This shows that the model is not reacting to the increased pumping in the same way that the natural system would react. This discrepancy provides less confidence in the second transient-state simulation than with the steady-state simulation. Therefore, true steady-state conditions for the second transient simulation were not truly reached in three years.

Sensitivity Analysis

A sensitivity analysis of the model demonstrated that the two continuous, confining layers that blanket the valley may have an impact on the water levels in the confined aquifers. The greatest head changes during calibration were produced in the confined aquifers by altering the vertical hydraulic conductivity of the confining layers. Groundwater heads also remained constant whether the vertical hydraulic conductivity of the continuous confining layers was altered in the center of the valley or towards the valley margins. The model-computed water levels in the unconfined aquifer were more sensitive to the discharge head of the cells representing springs, rivers, and evapotranspiration than to the conductance or evapotranspiration rate of the respective cells.

The sensitivity analysis also showed that altering the vertical hydraulic conductivity of the principal aquifer produced minimal head changes. The remaining hydraulic properties, namely the specific storage, porosity, effective porosity, and the specific yield, did not produce any significant head changes.

Limitations

This model, as with other numerical models, cannot perfectly simulate the natural environment. The model is based upon simplifying assumptions, but within limits, this model can assist in better understanding the interactions between surface water and groundwater in the unconsolidated basin-fill deposits in Cache Valley.

The model was created to simulate groundwater and surface interactions in the unconsolidated basin-fill deposits in both the Idaho and Utah portions of Cache Valley. In order to simulate such a large area, the size of the cells that compose the model were required to be relatively large. The large size of the cells does not allow the model to provide fine resolution. Therefore, it is not recommended that this model be used for small-scale applications.

The recharge to the unconsolidated basin-fill deposits through infiltration of precipitation is not uniform throughout the valley. The amount of recharge to the unconsolidated basin-fill deposits in various areas throughout the valley is not known. Therefore, it was assumed that precipitation recharges the unconsolidated basin-fill deposits uniformly over the entire basin. The uniform application over the entire valley does not account for the differences in soil types or precipitation intensity throughout the valley.

The amount of recharge to the unconsolidated basin-fill deposits of Cache Valley was assumed to be 20 percent of the total precipitation that fell on Cache Valley. This percentage was obtained from published infiltration rates from studies of other basins in Utah due to the lack of infiltration data for Cache Valley. The actual average infiltration rate in Cache Valley could be greater or less than the rate used for this model. Altering this infiltration rate would have a direct impact on the cells that simulate recharge and discharge to and from the unconfined aquifer.

In order to simulate recharge to the unconsolidated basin-fill deposits of Cache Valley through unconsumed irrigation water, the model required canal conveyance efficiency and on-farm efficiency values. The values used for input in the model were average values taken from the U. S. Soil Conservation Service (1976). Recharge to the unconsolidated basin-fill deposits through unconsumed irrigation water was also assumed to be distributed evenly over the irrigated areas. This assumption was necessary due to insufficient land application data for the irrigated areas. All canals were assumed to have equal canal conveyance efficiencies. This assumption was also necessary due to the lack of detailed information on the canal systems within Cache Valley. The cells which simulate recharge and discharge from the unconfined aquifer would be directly impacted through modifying canal conveyance efficiencies, on-farm efficiency values, and/or the distribution of unconsumed irrigation water.

High, Maple, and Mink Creeks discharge the greatest amount of water to the unconsolidated basin-fill deposits of Cache Valley. Consequently, it was assumed that these were the only creeks in which recharge to these deposits occurs, and that recharge from other streams is small compared to these three streams. The results of the model are based upon the aforementioned assumption, while any change in this assumption would impact other sources of recharge and discharge to and from the unconfined aquifer as well as the amount of water recharging the groundwater system through rivers. However, this assumption seems reasonable, and any error produced is probably negligible.

In order to simulate recharge to or discharge from the groundwater system through streams or rivers, MODFLOW (McDonald and Harbaugh 1988) requires a conductance value for input. The conductance value for a river or stream requires the river bed thickness and hydraulic conductivity of the river bed in order to calculate the value. These values were assigned according to river bed type and the size of the river. These values were also assumed to be the same within the same river system. The amount of discharge to river cells was distributed uniformly within the same river system. The rate of discharge from cells representing discharge from the groundwater system through rivers would be impacted if the conductance of the river bed material was altered or if the distribution of the discharge was altered. This alteration may also affect other sources of recharge and discharge to and from the groundwater system. However, any error associated with these values is probably small because estimated and model-computed fluxes are nearly identical.

Very little information exists for the various springs found throughout the unconsolidated basin-fill deposits in Cache Valley. Detailed information such as head distribution around the spring, aquifer hydraulic conductivity near the spring, distribution of fill material, number and size of the drain-pipe openings, the amount of clogging materials, and the hydraulic conductivity of clogging materials is not available. Due to the lack of sufficient detailed information and in order to calculate the drain conductance for this model, the River Package conductance was substituted for the Drain Package conductance (McDonald and Harbaugh 1988). The hydraulic conductivity for the area around the drain was also assumed to be the same as the value used for the river bed material. The quantity of water that discharges from the groundwater system through springs in various areas throughout the valley is not known. Consequently, the model simulates uniform discharge from each of the cells that represent springs. The actual rate of discharge from each of the cells that represent discharge from the groundwater system through springs could be refined if detailed spring information was available.

Discharge from the groundwater system in Cache Valley through evapotranspiration was distributed uniformly over the areas designated by Bjorklund and McGreevy (1971) as areas of evapotranspiration. This was done because the amount of discharge from each of the various evapotranspiration areas has not been quantified. The extinction depth for evapotranspiration in each of the various evapotranspiration areas also is not known. As a result, the extinction depth for cells that represent discharge from the groundwater system through evapotranspiration was assumed to be the same in each of the various evapotranspiration areas. The results of the model are based upon the aforementioned assumptions. Any change in these assumptions would impact other sources of recharge and discharge to and from the unconfined aquifer as well as the amount of discharge through evapotranspiration.

The values for the hydraulic parameters of the unconsolidated basin-fill deposits in Cache Valley were based upon the various groundwater areas designated by Bjorklund and McGreevy (1971). This is a generalization of the detail that exists within these deposits. The accuracy of the model could be refined if detailed information concerning hydraulic parameters was available.

Suggestions for Further Work

As was mentioned previously, the two continuous confining layers appear to restrict groundwater flow between the unconfined aquifer and the lower confined aquifer. Altering the vertical hydraulic conductivity of these layers during the sensitivity analysis also had an impact on the water levels within the confined aquifers. During calibration, the vertical hydraulic conductivity values of the confining layers were altered within an acceptable range for unweathered marine clay (Freeze and Cherry 1979). The actual vertical hydraulic conductivity value is not known. It would be useful to determine this value because of its impact on the groundwater flow between the unconfined and lower confined aquifers, and on the water levels in the confined aquifers. Robinson (1999) (Plate 1) shows that the two continuous confining layers and the upper confined aquifer are very distinct in the area of T12N, R1E, Sections 9, 16, and 17. This would be an ideal location in which to determine the vertical hydraulic conductivity of the two continuous confining layers.

The accuracy of this model could be greatly improved by performing a comprehensive spring survey. It would be useful to determine discharge values for the springs in order to better simulate the correct amount of discharge from each of the various springs cells. Also, the chemistry of the water being discharged from various springs in the valley could be used to determine whether the spring discharge originates from the lower, principal aquifer, the unconfined aquifer, or both.

In addition to the aforementioned information that is needed to more accurately simulate spring discharge, detailed information such as head distribution around the spring, aquifer hydraulic conductivity near the spring, distribution of fill material, number and size of the drain-pipe openings, the amount of clogging materials, and the hydraulic conductivity of clogging materials would be very useful. This information would allow the model to simulate spring discharge using the Drain Package conductance rather than the River Package conductance (McDonald and Harbaugh 1988), which would improve the ability of the model to simulate spring discharge.

Robinson (1999) describes the upper and lower confining layers as "clay grading to silt, sand, and gravel near the valley margins." The recharge to the principal aquifer through infiltration of precipitation is most likely greater through the sand and gravel portions of the confining layers near the valley margin than through the clay and silt in the middle portion of the valley, particularly since there is an upward groundwater gradient in the middle of the valley. The amount of recharge to the principal aquifer near the valley margins is not known. Therefore, the amount of recharge contributed through precipitation to the various aquifers in the saturated, basin-fill deposits was divided evenly over the basin according to cell area. A comprehensive annual precipitation study to determine the amount of precipitation throughout various areas of the valley would improve the accuracy of the model.

Due to insufficient data, recharge to the groundwater system through unconsumed irrigation water was assumed to be distributed evenly over the irrigated areas. A comprehensive study of the amount of water used for irrigated farmland in various areas throughout the valley would greatly improve the accuracy of the model in simulating recharge through unconsumed irrigation water. This study would not only involve the amount of water applied to irrigated farmland, but it would also involve determining soil infiltration rates, on-farm efficiencies, and canal conveyance efficiencies for the various areas of irrigated farmland throughout the valley.

Evapotranspiration is another aspect of the model where information was limited. Bjorklund and McGreevy (1971) performed a study on evapotranspiration. Their study focused mainly on the areas where most of the evapotranspiration occurs, such as wet meadowlands in the lower parts of the valley where the potentiometric surface is above the land surface. They did, however, estimate evapotranspiration from irrigated land and urban areas. A comprehensive study directed towards determining the actual quantity of water that evaporates from urban areas and irrigated land would also improve the accuracy of the model.

During calibration, general head boundary cells were introduced along the margin of the model domain. General head boundary cells introduced in the unconfined aquifer discharged water from the groundwater system through subsurface outflow, while general head boundary cells introduced in the lower confined aquifer recharged the groundwater system through subsurface inflow. This was done to achieve model-computed heads that are comparable to observed heads. In order to achieve steady-state, the model computed 69 cubic feet per second (2.0 cubic meters per second) discharged from the unconfined aquifer and 63 cubic feet per second (1.8 cubic meters per second) recharged the lower confined aquifer. The actual quantity of groundwater inflow or outflow is not known. It would be useful to locate and quantify sources of subsurface inflow and outflow from the consolidated rocks that line the margins of the valley, but it probably is not possible to measure this. Geophysical surveys or geologic mapping may assist in determining the locations of subsurface inflow and outflow.

The model of Cache Valley created for this thesis is and will be used as a simulation model. The construction of an optimization model used in conjunction with this simulation model would be beneficial in further understanding the water resources of Cache Valley. Using this simulation model as the basis, an optimization model could be developed to help water users determine the optimal consumptive use of surface water and groundwater in the valley.

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APPENDIX

Solution of Recharge to the Groundwater System Through
Infiltration of Precipitation and Unconsumed Irrigation Water

A. Recharge through infiltration of precipitation:

$$R = .2\,(P)(A)$$

where

R is recharge to the groundwater system due to infiltration of precipitation, in ft^3/yr;
P is average annual precipitation rate, in ft/yr;
A is area of study area, in ft^2;

$$R = .2\,(1.2\ ft/yr)(1.8x10^{10}\ ft^2)$$

$$R = 4.4x10^9\ ft^3/yr = 140\ ft^3/s$$

B. Recharge due to infiltration of unconsumed irrigation water:

$$R_1 = \frac{(TD \text{ x } CE)}{IA} \text{ x } (1.0\text{-}OE)$$

where

R_1 is recharge rate for year of interest, in ft;
TD is the total amount of water diverted to irrigated farmland, in acre-ft;
CE is the canal conveyance efficiency, in decimal form;
IA is irrigation company service area, in acres;
OE is on-farm efficiency, in decimal form;

$$R_1 = \frac{(205{,}000 \text{ acre-ft x } .59)}{158{,}835 \text{ acres}} \text{ x } (1.0 - .55)$$

$$R_1 = .34\ ft/yr$$

$$R = R_1 \text{ x } (A_f)$$

where

R is the recharge over the area of the farmland, in ft^3/s;
R_1 is the recharge rate for year of interest, in ft;
A_f is the area of farmland served by the canals, in ft^2;

$$R = .34\ ft/yr\ (6.9\ x10^9\ ft^2)$$

$$R = 2.3 \text{ x } 10^9\ ft^3/yr = 75\ ft^3/s$$