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THE HYDROTHERMAL SYSTEM OF THE LOWER EAST RIFT
ZONE OF KILAUEA VOLCANO: CONCEPTUAL AND
NUMERICAL MODELS OF ENERGY AND SOLUTE TRANSPORT

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION
OF THE UNIVERSITY OF HAWAI'I IN PARTIAL FULFILLMENT OF
THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

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GEOLOGY AND GEOPHYSICS

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and support while I concentrated on school and put up with a lot of long hours and stress.
ABSTRACT

The Lower East Rift Zone (LERZ) of Kilauea Volcano on the island of Hawai'i contains dike-impounded and basal ground water impacted by a hydrothermal flow system. This study was conducted to gain a better understanding of the dynamics of the shallow hydrothermal system and to investigate the implications of geothermal development on the system. Two phases of study were initiated: (1) a conceptual model of the shallow ground-water system was formulated based on previously published data and a shallow-well monitoring program and (2) numerical models of the hydrologic system were created based on the conceptual model.

Mathematical methods to remove the effects of ocean tide and barometric influence from continuous water-level records are presented. A numerical model was used to simulated the tidal response measured in the shallow wells and the results suggest that the aquifer south of the rift zone and in the eastern portion of the rift zone has a hydraulic conductivity in the range of 1,000 to 6,000 meters per day. Temperature profiles collected in six shallow wells show a less-dense warm water plume floating on the shallow aquifer. Comparisons with previous temperature profiles suggest that the hydrothermal system has been at "steady-state" for at least the last 20 years.

Two numerical models, CFEST and SUTRA are discussed and the models are verified by comparison to analytical solutions. SUTRA is more robust and accurate than CFEST for energy-transport simulations and is better suited for use in this study.

SUTRA simulations are used to estimate the flux of hot water entering the shallow aquifer from a deeper geothermal system. Simulated temperature profiles in three wells

v
match favorably with observed temperature profiles and estimates of the flux entering the shallow system through at least three different fractures range between 2.3 and 11.4 cubic meters per day of 100 °C fluid. The implications of changes in the hydrothermal system are investigated as varying discharge rates are introduced into the models and the changes in steady-state conditions are recorded.

Regional models of energy transport using SUTRA and solute-transport using CFEST are used to investigate the flow patterns in the regional hydrologic setting.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGMENTS</td>
<td>iii</td>
</tr>
<tr>
<td>ABSTRACT</td>
<td>v</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td>ix</td>
</tr>
<tr>
<td>LIST OF FIGURES</td>
<td>x</td>
</tr>
<tr>
<td>1. INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>2. PREVIOUS STUDIES</td>
<td>6</td>
</tr>
<tr>
<td>2.1. Geology and Hydrology of the Kilauea East Rift Zone</td>
<td>6</td>
</tr>
<tr>
<td>2.2. Numerical Modeling of Hawai‘i Aquifers</td>
<td>8</td>
</tr>
<tr>
<td>2.3. Numerical Modeling of Hydrothermal Systems</td>
<td>11</td>
</tr>
<tr>
<td>3. KILAUEA EAST RIFT ZONE GEOLOGY</td>
<td>15</td>
</tr>
<tr>
<td>3.1. Structure</td>
<td>15</td>
</tr>
<tr>
<td>3.2. Stratigraphy</td>
<td>18</td>
</tr>
<tr>
<td>4. KILAUEA EAST RIFT ZONE HYDROLOGY</td>
<td>24</td>
</tr>
<tr>
<td>4.1. Hydrologic Framework</td>
<td>24</td>
</tr>
<tr>
<td>4.2. Continuous Shallow Well Monitoring</td>
<td>32</td>
</tr>
<tr>
<td>4.2.1. Aquifer Response to Ocean Level Changes</td>
<td>36</td>
</tr>
<tr>
<td>4.2.2. Discussion of Tidal Simulations</td>
<td>43</td>
</tr>
<tr>
<td>4.2.3. Long-Term Groundwater Characteristics</td>
<td>46</td>
</tr>
<tr>
<td>4.2.3.1. Shallow Wells</td>
<td>46</td>
</tr>
<tr>
<td>4.2.3.2. Deep Wells</td>
<td>58</td>
</tr>
<tr>
<td>4.3. Thermal Characteristics of the Shallow System</td>
<td>60</td>
</tr>
<tr>
<td>4.3.1. Shallow Well Temperature Profiles</td>
<td>61</td>
</tr>
<tr>
<td>4.3.2. Discussion of Temperature Profiles</td>
<td>62</td>
</tr>
<tr>
<td>4.4. Chemical Characteristics of the Shallow System</td>
<td>68</td>
</tr>
<tr>
<td>5. COUPLED FLOW, ENERGY AND SOLUTE TRANSPORT MODELING</td>
<td>75</td>
</tr>
<tr>
<td>5.1. Mathematical Formulation</td>
<td>75</td>
</tr>
<tr>
<td>5.1.1. Flow Equation</td>
<td>76</td>
</tr>
<tr>
<td>5.1.2. Solute Transport Equation</td>
<td>78</td>
</tr>
<tr>
<td>5.1.3. Energy Transport Equation</td>
<td>81</td>
</tr>
<tr>
<td>5.2. CFEST Numerical Model</td>
<td>84</td>
</tr>
<tr>
<td>5.2.1. Features and Capabilities</td>
<td>85</td>
</tr>
<tr>
<td>5.2.2. Code Structure</td>
<td>88</td>
</tr>
<tr>
<td>5.2.3. Code Modifications</td>
<td>91</td>
</tr>
<tr>
<td>5.3. SUTRA Numerical Model</td>
<td>93</td>
</tr>
<tr>
<td>5.3.1. Features and Capabilities</td>
<td>93</td>
</tr>
<tr>
<td>5.3.2. Code Structure</td>
<td>95</td>
</tr>
<tr>
<td>5.3.3. Code Modifications</td>
<td>95</td>
</tr>
</tbody>
</table>
# LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>Table of previous Hawai‘i modeling studies</td>
</tr>
<tr>
<td>2.2</td>
<td>Table of previous hydrothermal modeling studies</td>
</tr>
<tr>
<td>4.1</td>
<td>Table of ERZ shallow well information</td>
</tr>
<tr>
<td>4.2</td>
<td>Table of tidal components</td>
</tr>
<tr>
<td>4.3</td>
<td>Table of tidal response in ERZ wells</td>
</tr>
<tr>
<td>4.4</td>
<td>Table of Green Lake temperature and conductivity profile data</td>
</tr>
<tr>
<td>5.1</td>
<td>Table of energy-transport model verification scenarios</td>
</tr>
<tr>
<td>6.1</td>
<td>Table of input parameters for model simulations</td>
</tr>
<tr>
<td>6.2</td>
<td>Table of discharging fracture energy-transport scenarios</td>
</tr>
<tr>
<td>6.3</td>
<td>Table of discharging fracture solute-transport scenarios</td>
</tr>
</tbody>
</table>
### LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Map of Hawai‘i showing location of study area</td>
<td>2</td>
</tr>
<tr>
<td>1.2</td>
<td>Map of Kilauea showing surface features</td>
<td>3</td>
</tr>
<tr>
<td>3.1</td>
<td>Map of Kilauea East Rift Zone</td>
<td>16</td>
</tr>
<tr>
<td>3.2</td>
<td>Map of ERZ surface deposits showing age</td>
<td>20</td>
</tr>
<tr>
<td>3.3</td>
<td>Map of ERZ surface deposits showing deposition methods</td>
<td>21</td>
</tr>
<tr>
<td>3.4</td>
<td>Map of ERZ surface deposits showing deposit morphology</td>
<td>22</td>
</tr>
<tr>
<td>4.1</td>
<td>Map of ERZ recharge estimates</td>
<td>25</td>
</tr>
<tr>
<td>4.2</td>
<td>Schematic of aquifer and freshwater lens</td>
<td>26</td>
</tr>
<tr>
<td>4.3</td>
<td>Map showing well locations and water levels</td>
<td>28</td>
</tr>
<tr>
<td>4.4</td>
<td>Plot of raw and filtered shallow-well water-level records</td>
<td>34</td>
</tr>
<tr>
<td>4.5</td>
<td>Plot of modeled and observed tidal response at Allison, Malama Ki and Kapoho Airstrip Wells</td>
<td>42</td>
</tr>
<tr>
<td>4.6</td>
<td>Map of tidal response in ERZ wells</td>
<td>44</td>
</tr>
<tr>
<td>4.7</td>
<td>Plot of barometric pressure, ocean-level and smoothed shallow-well water-level records</td>
<td>47</td>
</tr>
<tr>
<td>4.8</td>
<td>Plot of Allison Well analysis</td>
<td>51</td>
</tr>
<tr>
<td>4.9</td>
<td>Plot of Kapoho Airstrip Well analysis</td>
<td>52</td>
</tr>
<tr>
<td>4.10</td>
<td>Plot of Malama Ki Well analysis</td>
<td>53</td>
</tr>
<tr>
<td>4.11</td>
<td>Plot of Paradise Park Well analysis</td>
<td>54</td>
</tr>
<tr>
<td>Section</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>---------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>4.12</td>
<td>Plot of rainfall and filtered shallow-well water-level records</td>
<td>55</td>
</tr>
<tr>
<td>4.13</td>
<td>Plot of rainfall and filtered SOH water-level records</td>
<td>59</td>
</tr>
<tr>
<td>4.14</td>
<td>Plot of temperature profiles in ERZ shallow wells</td>
<td>63</td>
</tr>
<tr>
<td>4.15</td>
<td>Temperature profiles for GTW-3</td>
<td>64</td>
</tr>
<tr>
<td>4.16</td>
<td>Temperature profiles for Kapoho Airstrip Well</td>
<td>65</td>
</tr>
<tr>
<td>4.17</td>
<td>Temperature profiles for Malama Ki Well</td>
<td>66</td>
</tr>
<tr>
<td>5.1</td>
<td>Flowchart of CFEST model</td>
<td>90</td>
</tr>
<tr>
<td>5.2</td>
<td>Flowchart of SUTRA model</td>
<td>96</td>
</tr>
<tr>
<td>5.3</td>
<td>Plot showing CFEST match to Ogata-Banks analytical solution</td>
<td>99</td>
</tr>
<tr>
<td>5.4</td>
<td>Schematic showing shape of freshwater lens based on Ghyben-Herzberg</td>
<td>100</td>
</tr>
<tr>
<td>5.5</td>
<td>Plot showing CFEST match to isotropic aquifer analytical solution</td>
<td>103</td>
</tr>
<tr>
<td>5.6</td>
<td>Plot showing CFEST match to anisotropic aquifer analytical solution</td>
<td>105</td>
</tr>
<tr>
<td>5.7</td>
<td>Plot showing CFEST match to one-dimensional energy advection analytical</td>
<td>107</td>
</tr>
<tr>
<td></td>
<td>solution</td>
<td></td>
</tr>
<tr>
<td>5.8</td>
<td>Plot showing comparisons between CFEST and SUTRA, Runs V1-V4</td>
<td>111</td>
</tr>
<tr>
<td>5.9</td>
<td>Plot showing comparisons between CFEST and SUTRA, Runs V5-V8</td>
<td>115</td>
</tr>
<tr>
<td>6.1</td>
<td>Schematic of conceptual model of fracture discharge</td>
<td>119</td>
</tr>
<tr>
<td>6.2</td>
<td>Finite-element mesh used in discharging fracture simulations</td>
<td>121</td>
</tr>
<tr>
<td>6.3</td>
<td>Plot showing range of fluid density and fluid viscosity</td>
<td>124</td>
</tr>
<tr>
<td>6.4</td>
<td>Plot showing observed and simulated temperature profiles, GTW-3</td>
<td>127</td>
</tr>
<tr>
<td>Section</td>
<td>Description</td>
<td></td>
</tr>
<tr>
<td>---------</td>
<td>-------------</td>
<td></td>
</tr>
<tr>
<td>6.5</td>
<td>Plot showing observed and simulated temperature profiles, MW-2 .......... 130</td>
<td></td>
</tr>
<tr>
<td>6.6</td>
<td>Plot showing observed and simulated temperature profiles, Malama Ki Well ... 133</td>
<td></td>
</tr>
<tr>
<td>6.7</td>
<td>Plot showing simulated transient temperature changes in GTW-3, MW-2, and Malama Ki Well ................................................................. 137</td>
<td></td>
</tr>
<tr>
<td>6.8</td>
<td>Plot showing simulated temperature profiles after 40 years in GTW-3, MW-2, and Malama Ki Well ................................................................. 138</td>
<td></td>
</tr>
<tr>
<td>6.9</td>
<td>Plot showing simulated transient temperature changes after 3 years in GTW-3, MW-2, and Malama Ki Well ................................................................. 140</td>
<td></td>
</tr>
<tr>
<td>6.10</td>
<td>Plot showing simulated transient sulfate concentration changes for three-year simulations in MW-2 ................................................................. 144</td>
<td></td>
</tr>
<tr>
<td>6.11</td>
<td>Plot showing sensitivity analysis for discharging fracture simulations, transient results, A - K ................................................................. 146</td>
<td></td>
</tr>
<tr>
<td>6.12</td>
<td>Plot showing sensitivity analysis for discharging fracture simulations, temperature profile results ................................................................. 148</td>
<td></td>
</tr>
<tr>
<td>6.13</td>
<td>Finite-element mesh used in Paradise Park Well simulations .................. 157</td>
<td></td>
</tr>
<tr>
<td>6.14</td>
<td>Plot showing observed and simulated water levels in Paradise Park Well ...... 159</td>
<td></td>
</tr>
<tr>
<td>6.15</td>
<td>Plot showing observed and simulated water levels and temperatures in Malama Ki Well .................................................................................. 163</td>
<td></td>
</tr>
<tr>
<td>6.16</td>
<td>Plot showing observed and simulated water levels and temperatures in MW-2 .................................................................................. 166</td>
<td></td>
</tr>
<tr>
<td>6.17</td>
<td>Finite-element mesh used in CFEST regional simulations ........................ 174</td>
<td></td>
</tr>
<tr>
<td>6.18</td>
<td>Plot showing temperature distribution for CFEST regional simulations .......... 176</td>
<td></td>
</tr>
<tr>
<td>6.19</td>
<td>Finite-element mesh used in SUTRA regional discharging-fracture simulations .................................................................................. 177</td>
<td></td>
</tr>
<tr>
<td>6.20</td>
<td>Plot of regional temperature distribution for three fracture scenarios .......... 180</td>
<td></td>
</tr>
</tbody>
</table>
6.21 Plot of profiles and transient response for three fracture scenarios ............... 182

6.22 Plot of regional temperature profile used for CFEST simulations .................... 185

6.23 Plot of chloride distributions from regional CFEST scenarios -- first and second cases ........................................................................................................... 186

6.24 Plot of comparison of simulated head results, first and second case .................. 188

6.25 Plot of flow vectors and travel times to ocean for regional CFEST simulation, first case ............................................................................................................. 190

6.26 Plot of flow vectors and travel times to ocean for regional CFEST simulation, second case ..................................................................................................... 191

6.27 Plot of temperature distribution used for regional CFEST simulation, third case .............................................................................................................. 193

6.28 Plot of chloride distribution from regional CFEST scenario -- third case .......... 194

6.29 Plot of flow vectors and travel times to ocean for regional CFEST simulation, third case ..................................................................................................... 195
1. INTRODUCTION

The Lower East Rift Zone (LERZ) of Kilauea Volcano on the Island of Hawai‘i contains dike-impounded and basal ground water affected by a hydrothermal flow system (figs. 1.1 and 1.2). Currently, a 25 megawatt geothermal production facility in the LERZ taps a deep geothermal reservoir to produce electricity for the Island of Hawai‘i. This study, funded by the Hawai‘i Department of Business, Economic Development and Tourism (DBEDT), was conducted to determine the implications of geothermal development for the shallow ground-water system. Concerns about the vulnerability of the shallow aquifer to contamination and overdevelopment exist because, until this study was initiated, active monitoring of the shallow system was very limited. Specifically, some of the aspects which need to be addressed are:

1) the unknown effects to the shallow aquifer due to withdrawal of fluid from the deep geothermal reservoir;

2) a lack of knowledge about the degree of interaction between the shallow aquifer and the deep geothermal reservoir;

3) the effect that geothermal fluid withdrawal or reinjection would have on the ambient water quality (temperature and solute concentration) encountered in existing shallow wells and coastal springs;

4) how to differentiate between aquifer changes due to recharge and those caused by geothermal production;
Figure 1.1 Location of study area. (from Sorey, 1994)
Figure 1.2 Surface features of Kilauea Volcano.

(from Holcomb, 1987)
5) how quickly a geothermal fluid leak or spill could be detected in the current network of shallow wells and how far will the impacts be felt?

To assess the impacts of development and gain a better scientific understanding of the flow regime in the LERZ, two phases of this study were initiated. In the first phase, a conceptual model of the shallow ground-water system was formulated based on previously published data and a shallow-well monitoring program also funded by DBEDT. This report combines all of the existing geologic, hydrologic, and geochemical data collected in the past with data collected recently to present an up-to-date conceptualization of the ground-water flow in the LERZ and surrounding areas. The second phase of the study involved creating numerical models of the hydrologic system based on the conceptual model. This phase of the study involved modeling never previously attempted for Hawai‘i aquifers and was a major undertaking because of the complex nature of modeling both energy and solute transport in an aquifer containing both a rift zone and a basal flow system. These numerical models were used to further refine the conceptual model and as tools for evaluating the impacts of external influences on the hydrologic system.

The specific objectives of both the first and second phases of this report are summarized as follows:

1) present a review of the pertinent published literature pertaining to numerical simulation of ground-water flow in Hawai‘i aquifers, and hydrothermal modeling in general;

2) present a review of the available published information about KERZ geology and hydrology;
3) evaluate new hydrologic data collected as part of the State’s geothermal monitoring program;

4) refine the conceptual model of ground-water flow within and around the KERZ to provide a basis for numerical simulations of ground-water flow and transport;

5) construct a numerical model or models to simulate aspects of the KERZ flow system which incorporates new and existing data including recharge estimates, ground-water levels, temperatures, and solute concentrations.

6) verify that the models are accurate by comparing simulated results to results of test cases with known solutions;

7) validate the models by comparing simulated results of water level, water temperature, and solute concentration to results observed in the field; and

8) use the numerical models as tools in:

   a) investigating the effects of infiltrating recharge into the shallow aquifer;

   b) estimating information about the effects of fluid entering the shallow aquifer for the deep geothermal reservoir;

   c) predicting how quickly and to what extent potential geothermal-energy production-related changes to the thermal and solute input into the shallow aquifer could be monitored;

   d) predicting the impact of potential geothermal-energy production-related changes on existing wells and coastal springs.
2. PREVIOUS STUDIES

This section provides a review of all of the existing published data about the hydrogeology of the KERZ, a review of pertinent literature about modeling of Hawai‘i ground-water systems, and a brief review of published studies of hydrothermal modeling for systems located elsewhere in the world. The reviews are useful for demonstrating some of the features unique to this study.

2.1. Geology and Hydrology of the Kilauea East Rift Zone

As with most areas of Hawai‘i, the earliest studies of the geology and ground-water resources in the KERZ were performed by Stearns and Macdonald (1946). When geothermal development began to be considered in the 1970’s, several studies of geochemistry were carried out by Druecker and Fan (1976), McMurtry et al. (1977), and Kroopnick et al. (1978) among others. These studies concentrated on the results of sampling and analysis of wells and springs around the LERZ. Studies of the geochemistry continued through the 1980’s and early 1990’s and are still being performed (Janik et al. 1994; Thomas 1987, 1994; Tilling and Jones, 1991; Scholl et al. 1993, Novak, 1995). Each of these studies has added more to the understanding of the evolving conceptual model of the hydrothermal flow system, but at the same time each has presented new data which indicate there is still much to be explained.

Several recent or in-press publications have attempted to synthesize the sparse and sometimes unreliable data on ground-water levels, chemistry and temperatures to
present a conceptual model of the hydrology in the LERZ. Takasaki (1993) estimated recharge to the LERZ based on rainfall patterns and evapotranspiration. He estimated hydraulic conductivity values for the rift and flank areas based on his estimated recharge and hydraulic gradient measurements. He also suggested that ground water north of the LERZ flows to the east (parallel to the LERZ) and discharges at a high volume along the coast between Hilo and Cape Kumukahi. Ground water in the rift is thought to flow down rift to the east until it reaches the area east of the current geothermal production facility where fewer dikes are present to impound the water. It then flows in a direction more perpendicular to the topographic contours toward the coast between Cape Kumukahi and Opiehikao.

Ingebritsen and Scholl (1993) present a review and synthesis of existing data and make some suggestions regarding future investigations into the hydrothermal system. Some of the questions they feel need to be addressed involve the nature of the saltwater-freshwater interface in the rift zone, the effects of heat sources on the interface, and the causes of decreased permeability at depth in the rift zones. Sorey and Colvard (1994) present a similar review of existing data as part of an environmental impact statement and evaluate the potential for impact to the ground-water system from geothermal development. All of these studies are thorough but hampered by the scarcity of spatial and historical data. The most complete study of the ground-water geochemistry of the LERZ was recently completed by Novak (1995) as part of the same study as this project. Her report uses geochemical analyses from several ERZ area wells, as well as temperature, pressure and fluid conductivity data from continuous monitoring of shallow wells to characterize response to recharge events, develop transmissivity and hydraulic
conductivity values, delineate ground-water flow paths, establish the percentage of geothermal fluid mixed with ground water, and estimate the temperature and origin of thermal fluid in the shallow aquifer.

2.2. Numerical Modeling of Hawai‘i Aquifers

Interestingly enough, two of the earliest published numerical modeling studies pertaining to Hawaiian ground water were hydrothermal in nature (table 2.1). Murray (1974) performed a small-scale convection-cell finite-difference (F-D), vertical cross-sectional model to simulate the findings from the National Science Foundation (NSF) borehole drilled at Kilauea’s summit. His work suggested that the borehole was located about 200 m from the center of a proposed convection cell and fluid velocities were on the order of $4 \times 10^{-8}$ m/s. An impermeable zone of conductive heating 1,200 m below the surface was required to drive the convection cell. No efforts were made to describe regional flow patterns.

Cheng and Lau (1978) used a F-D, vertical cross-sectional model to simulate the development of a geothermal reservoir in an island aquifer bounded on both sides by a constant ocean boundary. This approach was semi-quantitative and involved a highly conceptualized aquifer with an impermeable caprock. Neither of these first models considered the effects of density due to solute concentration (i.e. freshwater/saltwater).

The earliest attempt at a regional scale flow model was performed for the Honolulu Aquifer Area 1 by Todd and Meyer (1970) and Meyer et al. (1974) at GE-TEMP0 for the
<table>
<thead>
<tr>
<th>Investigators</th>
<th>Study Area</th>
<th>Computer Model Used</th>
<th>Model Type</th>
<th>No. of Nodes or Elements</th>
<th>Site of Mesh</th>
<th>Includes Transition Zone?</th>
<th>Includes Thermal Effects?</th>
<th>Incorporates Rift Zone?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Murray (1974)</td>
<td>Kilauea Summit Area</td>
<td>unnamed convection cell model</td>
<td>2-D, cross-sectional, F-D, energy, transient</td>
<td>890 nodes</td>
<td>0.7 km x 0.7 km</td>
<td>NO</td>
<td>YES</td>
<td>YES</td>
</tr>
<tr>
<td>Cheng and Lau (1976)</td>
<td>general study of heat flow</td>
<td>unnamed</td>
<td>2-D, cross-sectional, F-D, energy, transient</td>
<td>3321 nodes</td>
<td>1.5 km x 1.5 km</td>
<td>NO</td>
<td>YES</td>
<td>NO</td>
</tr>
<tr>
<td>Wheatcraft (1979)</td>
<td>general study of injection wells</td>
<td>unnamed IAD model</td>
<td>2-D, cross-sectional, F-D, solute, transient</td>
<td>210 nodes</td>
<td>1 km x 0.1 km</td>
<td>YES</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Liu et al. (1983)</td>
<td>Pearl Harbor, Oahu</td>
<td>Tresscott model</td>
<td>2-D, areal, F-D, transient</td>
<td>285 elements</td>
<td>1.8 km x 1.9 km</td>
<td>NO</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Imaide (1984)</td>
<td>KERZ, Hawai‘i</td>
<td>Tresscott model, MOC</td>
<td>2-D, areal, F-D, solute, transient</td>
<td>60 elements</td>
<td>6 km x 4 km</td>
<td>NO</td>
<td>NO</td>
<td>YES</td>
</tr>
<tr>
<td>Eyre (1985)</td>
<td>Waikale - Hawai‘i Kai area, Oahu</td>
<td>AQUIFEM</td>
<td>2-D, areal, F-E, transient</td>
<td>~ 300 nodes</td>
<td>13 km x 8 km</td>
<td>NO</td>
<td>NO</td>
<td>YES</td>
</tr>
<tr>
<td>On and Lau (1987)</td>
<td>Pearl Harbor Aquifer, Oahu</td>
<td>Tresscott model, MOC</td>
<td>2-D, areal, F-D, solute, transient</td>
<td>400 elements</td>
<td>30 km x 20 km</td>
<td>NO</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Souza and Voss (1987)</td>
<td>Southern Oahu</td>
<td>SUTRA</td>
<td>2-D, cross-sectional, F-E, solute, transient</td>
<td>702 nodes</td>
<td>18 km x 1.8 km</td>
<td>YES</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Underwood and Shade (in press)</td>
<td>Kohala, Hawai‘i</td>
<td>SUTRA</td>
<td>2-D, cross-sectional, F-E, solute, transient</td>
<td>1281 elements</td>
<td>12 km x 1 km</td>
<td>YES</td>
<td>NO</td>
<td>YES</td>
</tr>
<tr>
<td>Underwood and Shade (in press)</td>
<td>Kohala, Hawai‘i</td>
<td>AQUIFEM-SALT</td>
<td>2-D, areal, F-E, transient</td>
<td>275 elements</td>
<td>28 km x 10 km</td>
<td>NO</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Oh (in progress)</td>
<td>Central Oahu</td>
<td>SHARP</td>
<td>2-D, areal, F-D, transient</td>
<td>3330 elements</td>
<td>7.5% of Oahu</td>
<td>NO</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>This Study</td>
<td>KERZ, Hawai‘i</td>
<td>CFEST</td>
<td>2-D, cross-sectional, F-E, solute, energy, transient</td>
<td>2698 elements</td>
<td>10.5 km x 0.6 km</td>
<td>YES</td>
<td>YES</td>
<td>YES</td>
</tr>
</tbody>
</table>
Department of the Interior. The first investigator in Hawai‘i who considered solute transport was Wheatcraft (1979), who performed a study of injection-well flow dynamics in a typical island aquifer which considered flow driven by the density differences between freshwater and saltwater. Liu et al. (1983) and Eyre (1985) used simple flow models which estimated the thickness of the basal freshwater lens based on the 40:1 Ghyben-Herzberg ratio ("sharp interface model") to evaluate ground water development situations on Oahu.

Imada (1984) combined a sharp interface flow model with the solute-transport method of characteristics (MOC) to simulate the head patterns in the LERZ and the flow of highly saline water around the HGP-A site. This model was two-dimensional and areal in extent, so heads and concentrations were vertically averaged. The effects of temperature were incorporated by varying fluid viscosity; however density effects were ignored. The aquifer was considered isotropic and the system was assumed to be at steady-state. In another study using MOC, Orr and Lau (1987, 1988) combined MOC and a mixing-cell model to model the transport of DBCP in the Pearl Harbor Aquifer.

Souza and Voss (1987), using the F-E, vertical, cross-sectional model SUTRA, were the first investigators to model the complete density-dependent freshwater/saltwater transition zone flow system. Their model through southern Oahu included the high permeability lava aquifer and the the low permeability caprock which acts as a confining unit. They used their model to investigate what parameters control the depth and thickness of the transition zone. Some of their findings were; 1) steady-state aquifer behavior is controlled by the horizontal conductivity of the basalt aquifer and recharge; 2) short-term transient hydraulic behavior is controlled by various parameters, mainly
specific yield, aquifer matrix compressibility, and vertical hydraulic conductivity of the basalt aquifer; 3) the regional distribution of concentration is insensitive to longitudinal dispersivity but highly sensitive to transverse dispersivity (Souza and Voss, 1987). A subsequent publication (Voss and Souza, [in press]) continues the investigation and presents findings on the effects of varying dispersivities depending on the ground-water flow direction. For both longitudinal and transverse dispersivity, a maximum and minimum dispersivity value is chosen to more realistically account for the anisotropy of the basalt aquifer.

Several U.S. Geological Survey (USGS) investigations are in press or underway including work in North Kohala by Meyer and Souza using SUTRA and AQUIFEM-SALT, and Central Oahu by Oki using SHARP. Preliminary flow models on Maui, Lanai, and Kauai are also being completed. All of the above-mentioned regional flow models assume an isothermal fluid and therefore density effects due to temperature are not considered.

2.3. Numerical Modeling of Hydrothermal Systems

Studies of energy and fluid transport in porous media using analytical and numerical models are numerous. For the purpose of this report, the following chronological review includes only those studies which discuss numerical modeling of actual large-scale single-phase hydrothermal flow systems (table 2.2). The earliest study was completed by Mercer et al. (1975) for the Wairakei hydrothermal field of New Zealand. They used a Galerkin-FE model with a two-dimensional areal mesh to match temperature and pressure measurements taken from wells. The major drawback to this
<table>
<thead>
<tr>
<th>Investigators</th>
<th>Study Area</th>
<th>Computer Model Used</th>
<th>Model Type</th>
<th>No. of Nodes or Elements</th>
<th>Size of Mesh</th>
<th>Includes Solute Transport?</th>
<th>Includes Free-Surface?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mercer et al. (1975)</td>
<td>Wairakei, New Zealand</td>
<td>unnamed</td>
<td>2-D, areal, F-E, single-phase, transient</td>
<td>41 elements</td>
<td>12 km x 5 km</td>
<td>NO</td>
<td>YES</td>
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<tr>
<td>Sorey (1976)</td>
<td>Long Valley, California</td>
<td>unnamed</td>
<td>3-D, F-D, single-phase, transient</td>
<td>410 nodes</td>
<td>35 km x 6 km</td>
<td>NO</td>
<td>YES</td>
</tr>
<tr>
<td>Fehn et al. (1983)</td>
<td>Galapagos Spreading Center</td>
<td>unnamed</td>
<td>2-D, cross-sectional, F-D, single-phase, transient</td>
<td>840 nodes</td>
<td>230 km x 5 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Ingebritsen and Sorey (1985)</td>
<td>Mt. Lassen, Oregon</td>
<td>unnamed model by Faust and Mercer</td>
<td>2-D, cross-sectional, F-D, 2-phase, transient</td>
<td>130 elements</td>
<td>10 km x 2.5 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Forster and Smith (1988)</td>
<td>general study of mountainous terrain</td>
<td>unnamed</td>
<td>2-D, cross-sectional, F-E, single-phase, steady-state</td>
<td>2200 elements</td>
<td>6 km x 2 km</td>
<td>NO</td>
<td>YES</td>
</tr>
<tr>
<td>Sammel et al. (1988)</td>
<td>Newberry Volcano, Oregon</td>
<td>PT</td>
<td>2-D, cross-sectional, F-D, single-phase, transient</td>
<td>459 nodes</td>
<td>40 km x 4 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Barroll and Reiter (1990)</td>
<td>Socorro, New Mexico</td>
<td>?</td>
<td>2-D, cross-sectional, F-D, single-phase, steady-state</td>
<td>?</td>
<td>14 km x 3 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Fisher et al. (1990)</td>
<td>Site 504, Mid-Ocean Ridge</td>
<td>PT</td>
<td>2-D, cross-sectional, F-D, single-phase, steady-state</td>
<td>825 elements</td>
<td>10 km x 1 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Fisher and Narasimhan (1991)</td>
<td>Guaynias Basin, Gulf of California</td>
<td>PT</td>
<td>2-D, cross-sectional, F-D, single-phase, steady-state</td>
<td>546 elements</td>
<td>2 km x 1 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Becker and Blackwell (1993)</td>
<td>Roosevelt Hot Springs, Utah</td>
<td>CCC</td>
<td>2-D, cross-sectional, F-D, single-phase, transient</td>
<td>287 elements</td>
<td>50 km x 6 km</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Sanford et al. (1994)</td>
<td>Poás Volcano, Costa Rica</td>
<td>HST3D</td>
<td>2-D, radial, cross-sectional, F-D, single-phase, transient</td>
<td>~ 1500 elements</td>
<td>6 km x 2 km</td>
<td>YES</td>
<td>YES</td>
</tr>
<tr>
<td>This Study</td>
<td>Kilauea ERZ, Hawai‘i</td>
<td>CFEST</td>
<td>2-D, cross-sectional, F-E, single-phase, transient</td>
<td>2698 elements</td>
<td>10.5 km x 0.65 km</td>
<td>YES</td>
<td>YES</td>
</tr>
</tbody>
</table>
approach was that vertical measurements had to be averaged throughout the entire thickness of the aquifer.

Sorey (1978) published a much-referenced and respected paper on numerical modeling of liquid geothermal systems and a companion paper (Sorey et al., 1978) on the hydrothermal system of Long Valley, California. His model was a three-dimensional, F-D model for single-phase flow in regional systems and contained equations of state for density and viscosity dependence on temperature.

Fehn et al. (1983) applied numerical modeling techniques to a study of the hydrothermal system at the Galapagos spreading center in the Eastern Pacific Ocean. Such a study can be useful because the concepts are essentially the same for fluid convection systems operating in the Earth's crust at mid-ocean ridges (MOR) and land-based hydrothermal systems. The MOR modeling studies are performed mainly to investigate heat flow and cooling patterns of the newly extruded basaltic crust and use vertical cross-sectional models perpendicular to the axis of the ridge.

Ingebritsen and Sorey (1985) used a two-dimensional, vertical cross-sectional model to simulate the hydrothermal system in the mountainous terrain of Lassen Volcanic National Park. Their cross-section included a constant-pressure surface which parallels existing topography and zones of higher permeability aquifer which concentrated fluid flow to the flanks of the mountain. A calibrated model was used to predict the effects of geothermal development on the flow system based on different withdrawal and reinjection scenarios.

Forster and Smith (1988) published a series of papers dealing with numerical modeling of hydrothermal systems in mountainous terrain. Although no site-specific
cases are presented, the results provide insight into the mechanisms which control
ground-water flow such as topographic slope, depth of circulation, and distribution of
recharge. In fact, it is the first regional study which included the effects of a free-surface
boundary condition based on recharge.

Other investigators who have done essentially the same type of two-dimensional,
vertical, single phase models include Sammel et al. (1988), Barrol and Reiter (1990),
Fisher et al. (1990), Fisher and Narasimhan (1991), and Becker and Blackwell (1993).
The only published study that involved both energy and solute transport is by Sanford et
al. (1994) who modeled the hydrothermal flow system of Poás Volcano, Costa Rica with
six radial cross-sections using HST3D (Kipp, 1986). They used their model to estimate
the travel time for heated brine to migrate from a summit crater lake to springs on the
flanks of the volcano. Their simulation included the effects of variable fluid density and
viscosity based on temperature and concentration changes but did not include a
freshwater/saltwater transition zone.

This study builds upon work modeling work performed both on Hawai‘i aquifers
and on work performed on other hydrothermal sites. It combines two density-dependent
processes, solute-transport modeling of the transition zone and energy-transport
modeling of the KERZ hydrothermal system. Previously reported transition-zone
modeling studies have only included isothermal conditions. Prior hydrothermal modeling
studies (with the exception of Sanford et al., 1994) have not included the effects of fluid
solute concentration on density. One of the other unique objectives of this study is to
provide insight into the way in which the effects of geothermal energy production can be
monitored using a shallow well network.
3. KILAUEA EAST RIFT ZONE GEOLOGY

This chapter provides a description of the structural and surficial geology of the KERZ based on an extensive amount of research that has been published for this area. The geologic information provides a framework on which the conceptual model of ground-water flow can be built.

3.1. Structure

Kilauea is an active tholeiitic shield volcano in the southeastern portion of the Island of Hawai‘i (fig.1.1). Eruptions occur from three principal structural features which are the summit caldera, the Southwest Rift Zone and the East Rift Zone (ERZ). The subaerial portion of the ERZ is about 55 km long; the submarine portion extends another 80 km as the Puna Ridge. The upper ERZ (UERZ) extends from Puhimau Crater to Mauna Ulu; the middle ERZ (MERZ) extends from Mauna Ulu to Heiheiahulu; and the lower ERZ (LERZ), from Heiheiahulu to Cape Kumukahi (figs. 1.2 and 3.1). The subaerial LERZ strikes N 65° E and is 3-4 km wide and about 23 km long. It is a constructional ridge, 50-150 m above the adjoining terrain, marked by low spatter ramparts and cones as high as 60 m (Moore, 1992). The altitude ranges from a high of 522 m at Heiheiahulu, a shield feature, to sea level at Cape Kumukahi and generally decreases from the southwest to northeast. Lava flows traveled away from the ridge to the northeast or southeast depending on the location of the source vent relative to the ridge.
Figure 3.1 Map of Kilauea East Rift Zone showing well locations and springs.
Traditionally it has been known that magma moves laterally into the ERZ through dikes from a reservoir beneath Kilauea's summit caldera. Recent findings based on widespread deformation at the summit and ERZ have led some researchers to suggest that both the summit and rift zones may be underlain by a thick, near vertical dike-like magma system at a depth of 3 to 9 km (Delaney et al., 1990). The implication is that magma is being supplied from a source rising from the base of the volcanic pile directly beneath the LERZ. Other evidence for this includes the electrical self-potential and seismic studies of Zablocki and Koyanagi (1979) and the petrologic studies by several investigators which found the vent deposits in the LERZ to be composed of more chemically evolved tholeiitic basalt. All of this data led Moore (1983) to suggest that the LERZ may have several long-lived secondary magma reservoirs whose associated hydrothermal-convection systems might be targets for geothermal exploration. Kauahikaua (1993), using drill hole data, estimated heat flow within the ERZ and summit to be between 370 and 820 mW/m² to a depth of 2 km.

Dike intrusions in the ERZ act as a wedge which pushes the south flank of the volcano towards the south leading to faulting, fracturing, and vertical and horizontal deformation. The concentration of dikes is high in the rift zone and the number is expected to decrease away from the rift zone to the NW and SE, although no deep drill holes are available to confirm this. Narrow grabens often form near eruptive and intrusive sites such as occurred during 1924 in Kapoho village when 3-4 m of subsidence occurred as the result of an intrusion. The existence of a transform fault perpendicular to the LERZ near Puulena Crater has been suggested by Zablocki and Koyanagi (1979) based on a northward shift of long eruptive fissures, a north-west trending electrical self-potential.
anomaly, and microearthquake epicenters along a northwest trend. Subsidence rates in the LERZ are estimated to be as high as 3.3 cm/year and extension rates are estimated to be about 2.4 cm/yr based on well records and leveling surveys (Delaney et al., 1993). The high concentration of intrusive tabular dikes and the faults and fractures associated with the continuous structural deformation are undoubtedly major controlling factors in the movement of groundwater in the ERZ, both vertically and horizontally.

The eruptive history of the ERZ has been determined from photogeology, extensive field mapping, oral tradition and carbon dating; several recent detailed reports and maps have been published (Holcomb, 1987, Moore and Trusdell, 1991, Moore, 1992, Moore and Trusdell, 1993). Holcomb (1987) reports a sympathetic behavior between the summit and rift zones over the last 500 years. Little flank activity has occurred when summit activity was sustained, and sustained summit activity has ceased when there was much activity in the rift zones. Five LERZ eruptions have been recorded in historic time; 1790, 1840, 1955, 1960, and 1961. The last three eruptions have lasted 88, 38, and 1 day, respectively. There have been no eruptions in the LERZ during the last 30 years.

3.2. Stratigraphy

Puna Basalt, covering all of the LERZ, is comprised of lava flows, vent deposits, and intrusions less than 10,000 years old. The oldest rock mapped is $2360 \pm 90$ years old (Moore and Trusdell, 1991) and the youngest was erupted in 1961. Seventy-five percent of the surface of the LERZ is covered by rocks less than 400 years old. According to Moore and Trusdell (1993) rocks below the surface occur as lava flows, dikes,
hyaloclastites, and pyroclastic deposits, and are equivalent in age to Hilina Basalt (30-100 ka). A database set for MapInfo, a Geographic Information System (GIS) software package was provided by Jim Kauahikaua of the Hawaiian Volcanoes Observatory and is useful for displaying the LERZ stratigraphy by different classifications. This GIS information is the same used by Moore and Trusdell (1991) for creating the USGS geologic map of the LERZ. Maps of the LERZ have been produced which show the distribution of rock units based on age (fig. 3.2), deposition mechanism (fig.3.3), and flow morphology (fig.3.4).

Moore (1992) and Moore and Trusdell (1993) provide a good description of the age and lithology of the LERZ deposits, and their reports are briefly summarized here. These descriptions cover only those rocks exposed at the surface but the subsurface rocks down to a least a depth of 1 km are expected to have the same characteristics. LERZ vent deposits include low (2-10 m) spatter ramparts, spatter cones as high as 60 m, two extensive lithic-rich deposits, local cinder-fall deposits, and the Heiheiahulu shield. Lava flows are aa or pahoehoe and typically extend 2-6 km from their vents; their range is a few tens of meters to more than 9 km. Flows range in thickness from 1-8 m at their margins to as much as 20 m where their central parts are exposed in sea cliffs.

Only a single small pahoehoe flow was mapped for the age range of 1,500-3,000 years old. There are 37 units for the range 750-1,500 years old, covering about five percent of the LERZ surface. Rocks include spatter deposits from 17 separate eruption vents, 11 flows associated with these vents, and 20 flows whose vents are covered by younger rocks. There are 54 units for the range 400-750 years old, covering about 20 percent of the study area. Of these units, 26 are spatter deposits formed during separate
Figure 3.2 East Rift Zone surface deposits - flow age.
Figure 3.3 East Rift Zone surface deposits - deposition methods.
Figure 3.4 East Rift Zone surface deposits - flow morphology.
eruptions, 18 lava flows associated with those spatter deposits, 27 flows whose vents are covered, and one tuff deposit. There are 15 map units for the range 200-400 years old and they cover about 20 percent of the study area. There are spatter deposits and lava flows of ten separate events, four lava flows whose vents are buried or outside the study area and one tuff cone. Twenty-five percent of the study area is covered by five deposits less than 200 years old. These units consist of vent deposits and associated lava flows and several small littoral cones at the coast.

The two deposits of ash and tuff are notable because they resulted from the interaction of magma with perched ground water, seawater, or a hydrothermal convection system. A deposit of unconsolidated, unstratified, lithic-rich tuff up to 20 m thick near Puulena Crater in the central LERZ was produced during violent phreatic and phreato-magmatic eruptions that probably resulted when ascending magma encountered perched ground water or a hydrothermal reservoir (fig. 3.3). Collapse near the end of the phreatic explosions formed the three largest pit craters, including Puulena Crater, on the LERZ. The deposit covers at least 4 km$^2$ and undoubtedly has been covered by younger lava over a much larger area. A weakly consolidated tuff deposit covers about 6 km$^2$ in the vicinity of Kapoho Crater. The base of the crater is only 20 m above sea level and violent interactions between magma and seawater undoubtedly produced the tuff.
4. KILAUEA EAST RIFT ZONE HYDROLOGY

This chapter discusses the existing literature about the hydrology of the KERZ and provides an analysis of new data collected as part of the State's geothermal monitoring program. The new data that has been collected includes continuous records of water level, water temperature and water specific conductivity in five shallow ERZ wells and temperature profiles from six shallow wells. Both the thermal and chemical characteristics of the shallow ground-water system are evaluated.

4.1. Hydrologic Framework

The abundant ground water in the East Rift Zone of Kilauea volcano is the result of high rainfall amounts on Kilauea and to some extent Mauna Loa. The ERZ lies in an area where rainfall averages about 254-320 cm/yr and pan evaporation averages about 130-190 cm/yr (Takasaki, 1993). Estimates of recharge based on these figures show the ERZ receiving anywhere from 25 to 250 cm/yr depending on location (fig. 4.1). This translates to about 0.87 million cubic meters per day (230 Mgal/d) of recharge within the boundaries of the ERZ.

Ground water in the vicinity of Kilauea's ERZ occurs in dike-impounded systems within the rift zone, a basal system in the northern and southern flanks and locally as a perched layer at Kapoho Crater (fig. 4.2). At the summit of Kilauea, ground water is impounded to an altitude of 611 m in the NSF-funded drill hole (Keller et al., 1979). Geophysical resistivity surveys also show high-level impounded water beneath the
Figure 4.1 East Rift Zone recharge estimates. (from Takasaki, 1993)
Figure 4.2 Schematic of aquifer and freshwater lens
summit and along the UERZ at least 16 km from the summit (Jackson and Kauahikaua, 1987a and 1987b). Impounded ground water was encountered near the MERZ-LERZ boundary during drilling of the MERZ-1 geothermal well (~ 160 m), drilled by True Geothermal Energy Co. and SOH-4 (44-260 m) although actual measurements were not made in MERZ-1 (Sorey and Colvard, 1994). Further down rift, ground water is impounded at lower elevations as the frequency of dikes in the aquifer probably decreases. Shallow water levels in the vicinity of the current geothermal development range between altitudes of 3-5 m (fig. 4.3).

Ground water both north and south of the ERZ occurs as a basal ground-water system of freshwater floating on saltwater separated by a transition zone of mixing. The typical Hawaiian basal ground water system has been discussed extensively in the published literature (for example, see Visher and Mink, 1964, Souza and Voss, 1987) and detailed descriptions of the hydrologic processes involved can be found there. Water level altitudes in wells south of the rift are about 1 m and the gradient is about 0.25 m/km toward the coast. The basal water discharges at the coast in the form of freshwater springs (fig. 4.3). The volume of basal discharge has not been measured, but estimates of discharge along the northeast coastline is estimated to be several million gallons per day (Druecker and Fan, 1976). Thermal-infra-red imagery (Fischer et al., 1966) also confirms the existence of diffuse basal discharge into the ocean where springs are not readily detectable.

Perched ground water is thought to occur at Kapoho Crater where the Kapoho Shaft Well collects water from a linear shaft located in a thick tuff layer. Green Lake, a 5-m deep body of water inside Kapoho Crater, also appears to be perched on low permeability
ash and the water in the lake is collected mainly from infiltration into the deposits of the cone.

Ground-water use in the ERZ is very limited; combined pumping averages about 50 l/s (800 gpm) (Sorey and Colvard, 1994). Of the seven large-capacity wells serving the area, only the Kapoho Shaft Well is located in the LERZ and it has been taken off-line recently. Rainfall catchment provides the main water supply for areas not connected to the County water supply system. The current geothermal production facility is air cooled and involves no consumptive use of ground water for producing power. The operation currently withdraws water from MW-1 and MW-3 only for minor use in the daily plant maintenance. The current PGV operating permit calls for a water supply rate of up to 32 l/s (500 gpm) to dissolve and inject produced gases in the event of an entire plant shutdown (Sorey and Colvard, 1994). MW-3 was flow tested for 32 hours at a rate of over 63 l/s with no more than about 15 cm of drawdown (written commun., PGV report to DLNR dated July 16, 1991) so it is capable of meeting the demands of an extreme situation.

The permeability of rocks in the ERZ probably ranges from greater than $10^{-10} \text{ m}^2$ to less than $10^{-15} \text{ m}^2$ depending on the morphology of the lava flows, the frequency of dike intrusion, and the degree of hydrothermal alteration (Ingebritsen and Scholl, 1993). Measurements of permeability in cores samples from the drill hole at the summit of Kilauea and the crust of the Kilauea Iki lava lake are $\sim 5 \times 10^{-17} \text{ m}^2$ and $3 \times 10^{-13} \text{ m}^2$ respectively, but these are representative of only small samples of the aquifer rock and are probably much denser than rift zone rocks. An estimate of the permeability of the summit rocks based on mud loss during drilling by Murray (1974) is $1 \times 10^{-14} \text{ m}^2$. Takasaki (1993) estimated hydraulic conductivity based on estimates of ground-water flow volume and hydraulic gradient. The equivalent permeability values are $1.2 \times 10^{-11} \text{ m}^2$ inside the
upper rift zone, $2.6 \times 10^{-10} \text{ m}^2$ inside the lower rift zone near the coast, $2.4 \times 10^{-10} \text{ m}^2$ south of the rift zone, and $2.3 \times 10^{-10} \text{ m}^2$ north of the rift zone.

No reliable measurements of permeability or hydraulic conductivity are available from aquifer tests in the ERZ or vicinity. An extensive literature review of the State of Hawai‘i Department of Land and Natural Resources (DLNR) files turned up reports of six attempts at conducting aquifer tests on shallow ERZ wells but none of the results were useful. Malama Ki Well was flow-tested for 23 hours at rates from 100-480 gpm, producing a maximum of about 0.8 ft of drawdown (DLNR, 1962), but the variable flow rates invalidates analysis of the data using analytical methods. Pulama Well was flow tested for 36 hours at rates ranging from 50-450 gpm but the drawdown was masked by tidal variations in the well (DLNR, 1964). Kapoho Shaft Well was flow tested for 48 hours at 360-400 gpm and 5.6 ft of drawdown was measured (written commun., Department of Water Supply, County of Hawaii). Analysis of these data are not useful because the well is an infiltration gallery with two 50-ft long sections. Paradise Park Well (Hawaiian Paradise #1) was flow tested for 24 hours at an unreported flow rate and no drawdown was reported (written commun., DLNR). Kapoho Airstrip Well was flow tested at 50 gpm; neither the duration nor drawdown of the test was reported (written commun., DLNR). The pumping test of MW-3 described previously was not useful because of the variable flow rates used in the test.

Estimates of permeability based on modeling studies are available for the ERZ and other Hawaiian aquifers. Imada (1984) reported permeability values of $4.5 \times 10^{-9} \text{ m}^2$ outside the rift zone, $9.8 \times 10^{-10} \text{ m}^2$ to rift-zone rocks within 10 km of Cape Kumukahi, and $9.8 \times 10^{-12} \text{ m}^2$ for the rift-zone rocks greater than 10 km from Cape Kumukahi. Other modeling studies for dike-free aquifers have produced permeability estimates within the
same range. Liu et al. (1983) modeled groundwater flow in the Pearl Harbor Aquifer and used permeability values of about $5 \times 10^{-10} \text{ m}^2$. Eyre (1985) was able to model groundwater flow in the Waialae and Wailupe-Hawaii Kai Koolau Basalt aquifers of southeastern Oahu using a permeability value of $2 \times 10^{-10} \text{ m}^2$. Souza and Voss (1987) used a value of $5.4 \times 10^{-9} \text{ m}^2$ for permeability of the basalt aquifer located in southern Oahu. Vertical permeability of the aquifer is usually estimated to be from one to three orders of magnitude less than the horizontal permeability due to layering of the lava flows.

Permeability in the ERZ appears to decrease with depth due to a higher concentration of dike intrusions and hydrothermal alteration which acts to fill openings by secondary mineralization. The presence of anhydrite, deposited as a vein-filling mineral, and of intermittent zones of intensely altered but low-permeability basalt in the cuttings from the deep geothermal wells on the LERZ tend to further substantiate this process (Thomas, 1987). Other vein-filling materials have been encountered in the HGP-A well, including chlorite and montmorillonite (Stone and Fan, 1978). A transition from advective to conductive energy-transport conditions occurs at depths varying from 1,100 to 1,400 m below sea level based on temperature profiles from the three deep Scientific Observation Hole (SOH) wells (Olson and Deymonaz, 1992). Ingebritsen and Scholl (1993) estimate that bulk permeability in that depth range (1,100 to 1,400 m below sea level) must be less than $10^{-14} \text{ m}^2$ or convection would still be occurring in this system but it is not clear to them whether pervasive intrusion or hydrothermal alteration is the major factor in causing such low permeability.
4.2. Continuous Shallow Well Monitoring

Table 4.1 lists information about the ERZ wells that were monitored with downhole instrumentation from 1992-94. The author performed downhole monitoring of hourly changes in water level, conductivity and/or temperature in seven shallow wells between May 1992 and November 1994 (Thomas, 1994). His report details numerous difficulties in collecting these data due to equipment failure, vandalism, etc. so the difficulties will not be discussed further here. Figure 4.4a shows the reliable raw water-level records that were obtained through this effort. All of the raw water-level records are plotted at calculated elevations above sea level based on the reported ground elevation at each well site. The most critical factor in making correct water-table elevation determinations is the exact elevation of the ground surface at each well. Normally, this value is easily determined through leveling surveys and remains constant over time. But as Delaney et al. (1993) have shown, the ground elevation over all of Kilauea including the ERZ is continually changing. They show that the ground elevation at the site of the Malama Ki and Kapoho Shaft Wells has been decreasing between 1983 and 1991 by about 2 and 3 cm/yr, respectively. Large earthquakes apparently can cause abrupt changes in these trends in either direction. The ground elevation at Malama Ki Well was reported in 1962 as 274 ft (83.5 m). Assuming the same subsidence rate (2 cm/yr) for the entire period from 1962 to 1994, the ground elevation may have moved by as much as 0.6 m. This is significant because the water level in Malama Ki Well is only a little more than 1 m above sea level. No attempt has been made to adjust any of the water levels to account for subsidence or rise of the ground surface because accurate level information for each well is not available at this time.
Table 4.1 ERZ shallow well information.

<table>
<thead>
<tr>
<th>Well #</th>
<th>Well Name</th>
<th>Major Use</th>
<th>Owner/User</th>
<th>Year Drilled</th>
<th>Latitude</th>
<th>Longitude</th>
<th>DEPTH</th>
<th>INCHES</th>
<th>FEET</th>
<th>ELEVATIONS IN FEET</th>
<th>REPRESENTATIVE MEASUREMENTS</th>
<th>PUMP TEST RESULTS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Temp</td>
<td>Cl</td>
</tr>
</tbody>
</table>

1. SOH wells completed with 7-inch casing to -2000 foot depth or smaller diameter liner from surface to total depth, slotted or perforated over the indicated elevation interval.
2. DOM=Domestic, GTH=Geothermal, IPR=Injection, MUN=Municipal, OB=Observation, OTH=Other, UNU=Unused.
3. HAWAII DWS=Hawaii Department of Water Supply, TMP=True Mid-Pacific, STATE DOHAWI=State Department of Water and Land Development, PGV=Puna Geothermal Ventures, HSCA=Hawaiian Shores Community Association.
4. Estimated from pressure survey run April 24, 1991 (approximately 5 months after completion).
5. Range of values measured during drilling. Lowest value applies to well depths of 2270-2700 ft. Highest values apply to depths less than 2270 ft.
6. Based on data in completion and monitoring reports, with reported depths to water reduced by 11 ft to account for difference in air-line depth listed on form and inferred from completion diagrams.
7. Maximum temperature encountered, or flowing temperature if followed by "F". Values in ( ) are representative of conditions in the depth interval 500-1000 ft.
8. CI value from bailed sample from approximately 600 ft depth, as listed in DLNR well file.
9. For HGP-A, range of values is for liquid separated at wellhead (corrected for steam flash) for samples collected over 1978-1989 production period. KS-1 produced only dry steam. KS-1A value is representative of wellhead samples collected during flow test in October, 1985 (corrected for steam flash). KS-2 value is for separated liquid collected during flow test in March 1991.

(from Sorey, 1994)
Figure 4.4 Raw and filtered shallow-well water-level records
The Hawaii Natural Energy Institute (HNEI) monitored pressure changes in the SOH wells starting in March 1992 using downhole pressure transducers. Exact water levels were never measured when the equipment was installed (H. Olsen, HNEI, oral commun., 1992) but approximate levels are available from the pressure data which has been converted to equivalent water-level data. The starting pressure on the transducer installed in SOH-1 was recorded as 1025 ± 10 psi. Assuming the borehole fluid is not saline and at a temperature of about 29 °C (Deymonaz, 1991), the average density of the fluid in the well above the transducer is about 996 kg/m³. The pressure on the transducer can be converted to an equivalent head in meters using the following equation:

\[
\frac{1,025 \text{ lb/in}^2 \times 6,897 \frac{\text{m}^2}{\text{in}^2}}{996.0 \frac{\text{kg}}{\text{m}^3} \times 9.81 \frac{\text{m}}{\text{s}^2}} = 724 \text{ m} \pm 7 \text{m}.
\]

Similar calculations are made for SOH-2 (392 m ± 4 m) and SOH-4 (410 m ± 4 m). The transducer in SOH-1 was positioned about 884 m below ground surface, which is at an altitude of about 183 m. The transducer in SOH-2 was 457 m below ground surface at an altitude of 91 m and the transducer in SOH-4 was 701 m below ground surface at an altitude of 366 m. Therefore the approximate water levels at the starting pressures for each transducer are:

- SOH-1: 183 m - 884 m + 724 m = 23 m ± 7 m,
- SOH-2: 91 m - 457 m + 392 m = 26 m ± 4 m,
- SOH-4: 366 m - 701 m + 410 m = 75 m ± 4 m.

Both Thomas' shallow and the HNEI deep well records show a high frequency component superimposed on longer period water-level fluctuations. In order to separate
the high and low frequency components, each record was filtered with a Fortran program
(lowpassvf) provided by Evelyn Roeloffs of the USGS Cascades Volcano Observatory (E.
Roeloffs, written commun.). Each record was separated with a high-pass filter into two
signals, one containing signals with periods less than 28 hours and the other with signals
having a period greater than 28 hours. Figure 4.4b,c shows the filtered signals for the
shallow wells. The high-frequency signals are the response of the water level in the
aquifer to the daily changes in ocean level due to Moon and Earth tides. A detailed
discussion of aquifer tidal response is provided in Chapter 4.2.1. The low-frequency
signals are the response of the water level in the aquifer to changes in recharge due to
seasonal precipitation patterns. A detailed look at these changes is provided in Chapter
4.2.3.

4.2.1. Aquifer Response to Ocean Level Changes

The water-level records containing the high-frequency ocean tidal response are
useful for determining some properties of the aquifer which is comprised of the ERZ and
flank basalts. Ferris (1951) presents the following analytical solution relating the tidal
response in wells to the hydraulic conductivity and storage coefficient of a confined aquifer
in direct contact with the ocean:

\[ h = h_o e^{-x \sqrt{\frac{z^2}{t_o^2}}} \sin \left( \frac{2 \pi t}{t_o} - x \sqrt{\frac{\pi S}{t_o T}} \right) \]  (4.1)
where:
- \( h\): ground-water head relative to mean sea level [L]
- \( h_o\): amplitude of tidal oscillation [L]
- \( x\): distance from coast [L]
- \( S\): storage coefficient [dim]
- \( t_o\): period of tidal oscillation [T]
- \( T\): aquifer transmissivity [L²/T]

The solution to eq. 4.1 shows that the tidal oscillations remain sinusoidal with a time lag and an exponential decrease in amplitude with distance from the sea, as follows,

\[
TE = e^{-\frac{x}{\sqrt{\frac{t_o S}{4\pi T}}}}, \quad LAG = x\sqrt{\frac{t_o S}{4\pi T}}
\]  

(4.2)

where:
- \( TE\): tidal efficiency, ratio of tidal amplitude in well to tidal amplitude in ocean
- \( LAG\): tidal lag, time between peak in ocean and peak in well

These equations can be used for unconfined aquifers if it is assumed (1) that vertical flow is negligible (Dupuit assumption), and (2) that variations in transmissivity resulting from fluctuations in the level of the phreatic surface are negligible (Erskine, 1991). The second assumption is satisfied, but the first is not in the area of the phreatic surface where fluid will be migrating vertically in and out of the pore space. The solution to this apparent problem is discussed later in this section.

There are many good articles in the literature discussing the details of spectral analysis of ocean and earth tides and the corresponding response which can be measured in wells (Williams and Liu, 1975, Hsieh et al., 1987, Rojstaczer and Riley, 1990, Ritzi et al. 1991, among others). The method used for analysis of the data in the ERZ was somewhat less rigorous than the methods presented in the above articles mainly because
the records were too short. The high-pass filtered water-level records were analyzed with a Fortran program also provided by Roeloffs (written commun., 1993) to decompose the various diurnal and semi-diurnal components which make up the signal. Table 4.2 shows the significant components of a real ocean tide.

<table>
<thead>
<tr>
<th>SYMBOL</th>
<th>PERIOD IN SOLAR HOURS</th>
<th>RELATIVE AMPLITUDE</th>
<th>DESCRIPTION OF COMPONENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>M2</td>
<td>12.42</td>
<td>1.000</td>
<td>MAIN LUNAR (SEMI-DIURNAL)</td>
</tr>
<tr>
<td>S2</td>
<td>12.00</td>
<td>0.466</td>
<td>MAIN SOLAR (SEMI-DIURNAL)</td>
</tr>
<tr>
<td>K2</td>
<td>11.97</td>
<td>0.013</td>
<td>SOLI-LUNAR DUE TO CHANGES IN DECLINATION OF SUN AND MOON</td>
</tr>
<tr>
<td>K1</td>
<td>23.93</td>
<td>0.584</td>
<td>SOLI-LUNAR</td>
</tr>
<tr>
<td>O1</td>
<td>25.82</td>
<td>0.415</td>
<td>MAIN LUNAR</td>
</tr>
<tr>
<td>P1</td>
<td>24.07</td>
<td>0.193</td>
<td>MAIN SOLAR</td>
</tr>
</tbody>
</table>

SOURCE: WILLIAMS AND LIU, 1975

The amplitude of the three diurnal components (K1, O1, and P1) and the three semi-diurnal components (M2, S2, and K2) for each well record were added to obtain one diurnal and one semi-diurnal amplitude for each well (table 4.3). The ocean level record from Hilo Harbor was obtained from the TOGA Sea Level Center at the University of Hawai‘i for the 1992-94 time period. The record was smoothed from the original six-minute sampling interval to hourly values and filtered with the high-pass filter. The high-frequency component was then analyzed in the same manner as the well records to obtain the amplitude of each of the six diurnal and semidiurnal tidal components (table 4.3). The diurnal and semi-diurnal tidal efficiency of each well was then calculated from
Table 4.3 Kilauea East Rift Zone monitoring-well tidal responses.

<table>
<thead>
<tr>
<th>WELL NAME</th>
<th>AMPLITUDE OF TIDAL COMPONENTS (M)</th>
<th>LAG (MINUTES)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>M2</td>
<td>S2</td>
</tr>
<tr>
<td>Hilo Harbor</td>
<td>0.185</td>
<td>0.083</td>
</tr>
<tr>
<td>Allison</td>
<td>0.097</td>
<td>0.048</td>
</tr>
<tr>
<td>Kapoho</td>
<td>0.017</td>
<td>0.014</td>
</tr>
<tr>
<td>Malama Ki</td>
<td>0.036</td>
<td>0.017</td>
</tr>
<tr>
<td>Paradise Park</td>
<td>0.001</td>
<td>0.001</td>
</tr>
<tr>
<td>SOH-1</td>
<td>&lt;0.000</td>
<td>&lt;0.000</td>
</tr>
<tr>
<td>SOH-2</td>
<td>0.009</td>
<td>0.004</td>
</tr>
<tr>
<td>SOH-4</td>
<td>&lt;0.000</td>
<td>&lt;0.000</td>
</tr>
</tbody>
</table>

WELLS WITH INSUFFICIENT RECORDS FOR SPECTRAL ANALYSIS

<table>
<thead>
<tr>
<th>EFFICIENCY</th>
<th>COMMENTS</th>
</tr>
</thead>
<tbody>
<tr>
<td>MW-2  ~3%</td>
<td>based on inspection of total signal</td>
</tr>
<tr>
<td>MW-3 &lt;1%</td>
<td>based on inspection of pumping test record</td>
</tr>
<tr>
<td>GTW-3 ~3%</td>
<td>based on inspection of 1 day signal</td>
</tr>
<tr>
<td>Kapoho Shaft ~12%</td>
<td>based on inspection of several day record</td>
</tr>
<tr>
<td>Green Lake ~12%</td>
<td>based on 1 day record</td>
</tr>
</tbody>
</table>
the ratio of the well amplitude to the ocean amplitude. In some cases the length of the record was not long enough to allow decomposition of the P1 component. For these records, only the K1 and O1 components were used in the efficiency calculation. Tidal efficiencies range from less than 1 percent to greater than 50 percent and are slightly higher for the diurnal tidal component. This is to be expected because the diurnal component travels at a lower frequency (i.e., longer wavelength) and hence penetrates more efficiently into the aquifer.

Tidal lags were determined in a simpler manner by averaging the time between maxima and minima in the wells and the corresponding maxima and minima in the Hilo ocean record. The lags were then shifted to account for the slight difference in high tide times between Hilo and Cape Kumukahi. High tide at South Point, Hawai'i is about 30 minutes later than at Hilo (oral commun., TOGA Sea Level Center), so fifteen minutes were subtracted from each estimated lag because Cape Kumukahi is about halfway between Hilo and South Point. For some of the wells, no lag was determined because the measured response was not large enough for an accurate determination.

As discussed above, the analytical equations for tidal lag and efficiency were derived for a confined aquifer case. To avoid the problems involved with using these equations, aquifer hydraulic conductivity and specific yield were estimated with the aid of a simple numerical ground-water-flow model. A rectangular, 11 x 201 node finite-element mesh was created and the flow characteristics were modeled with the numerical model SUTRA (Voss, 1984), which was modified for a water-table condition by Voss. The modification to the model allows for storage of fluid at the water table by assigning a value of specific yield to the top row of nodes in the mesh (Souza and Voss, 1987). The mesh
represented a cross-section through an aquifer 100 m deep and 10,000 m wide with one side in contact with the ocean. The top, bottom, and left side of the mesh were considered no-flow boundaries and the remaining side, representing the ocean was modeled as a specified pressure boundary of sinusoidally varying pressure. The pressure changes were equivalent to a 1-meter tidal range with a period of either 12 or 24 hours. Each simulation progressed for one tidal cycle (12 or 24 hours) and pressures were recorded at the top nodes of the mesh at various distances away from the ocean boundary (Table 6.1, discussed later, lists the parameters used in the simulation). The aquifer permeability and specific yield were varied systematically in an attempt to find the combination which provided the best fit to the measured tidal response in three shallow wells (Allison, Malama Ki, and Kapoho Airstrip) at their respective distances away from the coast. Increasing the aquifer thickness to greater than 100 m had no effect on the results.

Figure 4.5 shows the best-fitting model simulations for several different combinations of permeability and specific yield for a twelve-hour tidal cycle. The best fitting permeability values \(8.3 \times 10^{-10}\) to \(5.0 \times 10^{-9}\) m\(^2\) equate to a horizontal hydraulic conductivity in the range of 1,000 to 6,000 meters per day. These values compare favorably with other estimates of permeability in Hawai‘i basalt aquifers, which range from \(2 \times 10^{-10}\) to \(5.0 \times 10^{-9}\) m\(^2\) (Williams and Soroos, 1973, Imada, 1984, Eyre, 1985, Souza and Voss, 1987, Takasaki, 1993). The specific yield or "effective porosity" fits best in the range of 1 to 3 percent and are significantly lower than the usually stated porosity values of 5 to 10 percent for basalt. This may be explained by thinking of the specific yield as an "effective porosity" which only accounts for the interconnected pore space, the permeable pathways that transmit fluid through the rock matrix. Much of the total porosity may be
Figure 4.5 Modeled and observed tidal response at Allison, Malama Ki and Kapoho Airstrip Wells
saturated but consists of a "dead-end" pore space that does not effectively transmit fluid. Geophysical techniques used to estimate basalt porosity cannot differentiate between "effective" and total porosity.

Using tidal analysis to estimate aquifer permeability and specific yield is especially useful in Hawai'i's basalt aquifers where pumping tests typically produce ambiguous results due to the difficulty of creating significant aquifer drawdowns. The tidal method produces regional permeability estimates between the well and the coast, whereas pumping tests generally are only representative of the local conditions around a well.

4.2.2. Discussion of Tidal Simulations

Figure 4.6 shows the 12-hour and 24-hour tidal efficiencies for the monitored wells around the ERZ. Several features of this figure are worthy of further mention. Five wells (MW-2, MW-3, GTW-3, SOH-1, and SOH-4) inside the ERZ have much lower efficiencies than the flank wells outside the rift (Allison and Malama Ki). These low efficiency values indicate that the tidal signal is not propagating as effectively into this part of the aquifer. The most reasonable explanation for this is the presence of low permeability dikes between the ERZ wells and the flank wells. Model simulations predict that the efficiency in MW-2, located about 4,800 m from the coast, would be about 12-15 percent in a dike-free aquifer with a hydraulic conductivity around 4,000 m/d. But because the measured efficiency is only about 3 percent the model was changed to include a lower permeability zone representing the rift zone, which is a basalt aquifer containing many dikes. This edge of the zone was estimated based on surface features to be about
Figure 4.6 Observed tidal response in East Rift Zone wells.
midway between the Malama Ki Well and MW-2. The permeability of the rift zone in the model was lowered about two orders of magnitude below the flank permeability to produce acceptable efficiencies at the MW-2 location, 4,800 m from the coast. Although the Kapoho Airstrip Well is located in the center of the rift zone, the efficiency is not low like the other rift zone wells. This suggests that the pressure signal is traveling uprift from the coast at Cape Kumukahi and not encountering any low permeability dike zones at least as far uprift as the Kapoho Airstrip Well. The permeability in the rift zone for at least 4-5 km from the coast is the same as the permeability of the southern flank lavas perpendicular to the rift zone. The low efficiency in GTW-3 also suggests that low permeability dikes are encountered by the tidal signal somewhere between the GTW-3 location and the Kapoho Airstrip Well.

The other shallow well in the area that was monitored, Kapoho Shaft Well, has a tidal range of about 10 cm (12 percent efficiency). A crude measurement of water level was attempted in Green Lake, inside Kapoho Crater, and fluctuations due to ocean tides were found to also be about 10 cm although the data logger that was installed was somewhat suspect. These values are fairly low even though the both measuring sites are located close to the coast. It is believed that these low tidal responses are due to the fact that Green Lake, and to some extent the Kapoho Shaft Well, are perched above the basal lens by the Kapoho Cone Tuff, which would be expected to have a low permeability. There are many wells in Hawai‘i which draw water from aquifers perched above the basal lens on low permeability ash or tuff layers (see Stearns and MacDonald, 1946, pp 254-256, 263-269). The other shallow well for which tidal response data are available is the Hawaiian Paradise Park Well located about 15 km north of the ERZ in what would be
considered flank lava flows. The efficiency measured at this well suggests that the aquifer hydraulic conductivity is about 400 m/d. The cause of this relatively low value of hydraulic conductivity will be discussed in further detail in Chapter 6.2.3.

Tidal analysis was also carried out on continuous pressure records from the three deep SOH wells. The values for SOH-1 and SOH-4 are less than 1 percent, which is typical for wells located in the ERZ fairly distant from the coast. The efficiency at SOH-2 is fairly close to the efficiency of Kapoho Airstrip Well even though SOH-2 is open to the aquifer only below an altitude of -1,200 m (H. Olsen, HNEI, written commun., 1993). This indicates that the effects of the ocean tide are felt fairly deeply in the aquifer near the coast.

4.2.3. Long-Term Ground Water Characteristics

**Shallow Wells**

The low-pass filtered hourly water-level records (fig. 4.7c) reflect long-term changes in the level of water in the aquifer caused by long-term recharge patterns and by fluctuations in the ocean level. This is evident by comparing the low-pass filtered Hilo ocean level record (fig. 4.7b) with the water-level records. The most visible correlation is in the Kapoho Airstrip Well and Malama Ki Well records around July 1992 and mid-August 1993 where peaks in water level closely match peaks in the Hilo ocean level. The correlation between aquifer water level and ocean level is easily explained by picturing the
Figure 4.7 Barometric pressure, ocean level and East Rift Zone shallow well records.
lens of less-dense freshwater bobbing up and down on the underlying saltwater. The
cause of the ocean level changes appears to be corresponding changes in atmospheric
pressure. A rise in barometric pressure will cause a corresponding decrease in ocean
level as the "effective" weight of the column of air pressing down on the ocean surface is
increased. The ocean level can be looked at as an inverse barometer. The barometric
pressure record for the Old General Lyman Field Airport in Hilo was obtained from NOAA
(1994) for the time period May 1992 through March 1994. The hourly record of pressure
in millibars was converted to pressure in meters of head, filtered with the high-pass filter
and inverted to produce the record plotted in figure 4.7a. The same comparable peaks in
the water-level and ocean-level records can be seen in the inverted barometric pressure
record.

The changes in water level due to ocean changes ("atmospheric changes") are as
much as 0.1 m in a week. These changes are large enough to mask changes in the water
level due to recharge ("recharge changes") so it is important to remove the atmospheric
changes from the water-level records. The technique for removing atmospheric effects
from water-level records used in this study was described by Quilty and Roeloffs (1991).
The low-pass filtered well record and the ocean record for the same time period were
loaded into Matlab, a numerical computation software package on the SOEST computer
network. The well record and ocean record were then broken down into smaller sections
of equal length so an average estimate could be obtained. A spectral estimate of each
well-record section and corresponding ocean-record section was obtained with the
Fast-Fourier-Transform (FFT) algorithm;

\[
F_w(\omega) = \text{fft}(f_w(t))
\]  

(4.3)
where $f_w$ or $f_o$ denotes the original well or ocean record and $F_w$ or $F_o$ denotes the transformed record. The estimated frequency-domain auto-correlation of each ocean-record section;

$$G_{oo}(\omega) = (F_o(\omega) * F_o(\omega)),$$

and the estimated frequency-domain cross-correlation of each ocean-record section and each well-record section;

$$G_{ow}(\omega) = (F_o(\omega) * F_w(\omega))$$

were then obtained. The $*$ symbol in (4.5) and (4.6) represents convolution. The average auto-correlation and cross-correlation was then calculated by using all "reliable" individual section estimates. For each record, certain sections were not used because they provided poor cross-correlation records. This was usually because the water-level record contained frequencies not correlateable with the ocean record. A transform function for the whole well record was then computed with the following relationship:

$$H(\omega) = G_{owave}(\omega)/G_{o3ave}(\omega).$$

The transfer function, $H(\omega)$, is the mathematical representation of the effect of the aquifer on the signal of ocean level change as it propagates through the aquifer. The transfer function is then convolved with the spectral estimate of the ocean level record and this result is then operated on with the inverse FFT algorithm;

$$s(t) = \text{ifft}(H(\omega) * F_o(\omega)).$$

Finally, this result is subtracted from the well water-level record as follows,

$$r_w(t) = f_w(t) - s(t)$$
to obtain a water-level record ($r_w$) with no ocean-related atmospheric effects. This technique was used on the water-level records from Allison Well (fig. 4.8), Kapoho Airstrip Well (fig. 4.9), Malama Ki Well (fig. 4.10), and Paradise Park Wells (fig. 4.11). Figure 4.12 shows the resulting filtered water-level records for these four wells and MW-2 compared to daily rainfall records for three ERZ sites. The following discussion will focus on details of each of the five wells separately because each record has some unique features.

**Allison Well**

The Allison Well record is continuous for four months, November 1993 to March 1994 and is fairly smooth. The record remains steady over the four month period at about 1.7 m above sea level. The largest fluctuation (0.1 m) occurs at the beginning of December but it does not correlate very well with any recharge events.

**Kapoho Airstrip Well**

The Kapoho Airstrip Well record, which is in two sections of seven and nine months respectively, is the longest one collected. An additional step was necessary to produce the water-level record in figure 4.9. Upon inspection of the low-pass record, it is apparent that it contains a systematic rise in the data from about September to December 1992, which is thought to be due to a gradual equipment failure. After removing the atmospheric effects, the systematic rise was removed by subtracting a fitted exponential
Figure 4.8 Allison Well water-level analysis.
Figure 4.9 Kapoho Airstrip Well water-level analysis.
Figure 4.10 Malama Ki Well water-level analysis.
Figure 4.11 Paradise Park Well water-level analysis.
Figure 4.12 Final filtered water-level records for East Rift Zone shallow wells and rainfall records.
function from the last three and a half months of the record. This has flattened out the data
effectively and the result is a fairly smooth record of water level in the well. The record
suggests a gradual decline in water level in the aquifer from May to mid-November of 1992
and then a rise through December caused by the big rainfall events (> 60 cm in 24 hours
at Pahoa) right before the beginning of December. The record remains fairly flat from July
1993 through February 1994 but shows a slight rise in water level in August due to another
rainy period. The sudden drop in water level in February 1994 does not correlate with any
precipitation changes and is difficult to explain. One possible explanation is that the
fluctuation is earthquake related. On February 1, a deep earthquake of magnitude 5.2
occurred about 19 km beneath the summit of Kilauea. The USGS has recorded
earthquake-related water-level fluctuations at a well-instrumented site in Parkfield,
California (E. Roeloffs, USGS, oral commun.).

Malama Ki Well

The Malama Ki Well record has only one good continuous section from mid-May
to mid-August of 1992. The other sections are only about 20 days long each and the
reliability of the downhole sensors is somewhat questionable for these time periods. The
final water-level records are flat from May to October of 1992 and do not show much
water-level change due to recharge. Unfortunately, no record is available for the big
rainfall events at the end of November 1992. The short segments in April and May 1993
are not very smooth which is probably due to data filtering problems associated with
records that are not long enough for reliable analysis.
Because MW-2 is located inside the rift, atmospheric effects are damped and the low-pass record is fairly flat. For completeness, the transfer function for this well was calculated and applied to the MW-2 record, but the result was very noisy and much rougher than the input signal. Therefore, the filtering method was not used for estimating the final water-level signal. The signal shows a steady water-level decline of about 0.4 m from April to August 1993. A slight rise occurs at the end of the record about 20 days after a period of heavy recharge during the end of July. A lack of data makes it difficult to determine if this trend would have continued.

Paradise Park Well

The Paradise Park Well record is nearly continuous from October 1992 through September 1993. Unfortunately, the Hilo ocean level record is missing the months of January and February 1993 so a section of the well record could not be analyzed. An obvious difference between the Paradise Park Well record and the other well records is the large change in water level (~ 0.5 m) corresponding to the late-November 1992 recharge events. It appears that recharge is moving from the ground surface to the aquifer very quickly as a "slug" of water and causing a nearly instantaneous rise in the aquifer water level. The water level then declines smoothly over a period of about 10-15
days as the water flows away. This type of water-level response to rapid recharge suggests that the aquifer in the vicinity of the Paradise Park Well has a relatively high vertical permeability and a relatively low horizontal permeability. The vertical permeability must be high to allow the slug of recharge to reach the water table so quickly and the horizontal permeability must be relatively low to allow the level in the aquifer to rise so quickly and decline slowly. The low tidal efficiency (1 percent) for this well helps to back up this explanation. It is questionable whether the filtering method does a good job on this section of the well record because of the large fluctuations due to recharge. The filtered record seems noisier than the unfiltered record for the October-December 1992 and March-June 1993 sections of the record while the July-September, 1993 section appears to be cleaner. The filtering method is probably not successful for this record for the same reason it is inappropriate for the MW-2 record. Both wells appear to be located in sections of the aquifer which dampen out the ocean level signal almost completely. Therefore the transfer functions are not really meaningful representations of the effects of the aquifer on the signal.

Deep Wells

Figure 4.13 shows the low-pass records for the three deep wells that were monitored, SOH-1, 2, and 4. All three records show a steadily declining water level through December 1992 followed by a rise until March 1993 and then a decline through March 1994. The record for SOH-4 shows a fairly substantial change of 5 m in water level between December 1992 and February 1993. The other two wells have fluctuations of
Figure 4.13 Smoothed water-level records for SOH wells and rainfall records.
about 1 m for the same time period. This pattern indicates that the aquifer around SOH-4 is less permeable than around SOH-1 and SOH-2. This is probably due to an increase in the frequency of low-permeability dikes in the upper section of the rift zone. The rise in water level in December 1992 is about 10-15 days after the heavy rainfall events at the end of November. Close inspection of the records for SOH-1 and SOH-4 at February 1, 1994 reveals a slight rise in pressure in each well which may also be attributed to the magnitude 5.2 earthquake mentioned above.

4.3. Thermal Characteristics of the Shallow System

The first published mention of thermal water in the ERZ was in 1864 for the Blue Grotto, a 32°C pool of water near the base of Puu Kukae, a prehistoric cinder and spatter cone just east of the village of Kapoho (MacDonald, 1973). This pool was subsequently covered by the 1960 Kapoho eruption. The temperature of the warm springs at Pohoiki (Isaac Hale Spring) were measured by MacDonald in 1950 and reported to be about 33°C. Thermal infra-red imagery (Fischer et al., 1966) revealed the presence of more warm springs along the coast between Cape Kumukahi and Kamaili (fig. 4.3). Exploratory drilling in 1961-62 in the LERZ (GTW-1, -2, -3, and -4) and south of the rift zone (Malama Ki Well) revealed the presence of elevated temperature in the shallow aquifer. As interest in geothermal development grew, more shallow and deep wells were drilled and the distribution of warm water in the shallow aquifer was shown to be highly variable and localized but not without order. Wells drilled north of the ERZ have not encountered any water with temperatures indicating the effects of significant geothermal heating.
4.3.1. Shallow Well Temperature Profiles

On June 1 1994, temperature profiles were measured in six of the shallow ERZ wells by Elizabeth Novak, Harry Olsen (HNEI), and Stephen Gingerich using a Spafford probe provided by the HNEI. The probe contained a thermistor which was attached to a volt-meter to provide measurements of resistivity in the probe. The probe was lowered into the well using a hand-winチch and the depth was read from a mechanical depth-counter which provided readings in tenths of meters. The probe was calibrated by recording the resistivity readings at six known temperatures between 0°C and 100°C and fitting a fifth-order polynomial to the data. The operating manual for the probe provided the following equations for converting the resistivity measurements into temperatures:

\[
R = ALOG\left(\frac{400.0 \times KOHM}{400.0 - KOHM} - CABLE\right) \quad (4.10)
\]

\[
TEMP(°C) = \left(\frac{1}{T}\right) - 273.15 \quad (4.11)
\]

\[
T = C(0) + C(1) \times R + C(2) \times R^2 + C(3) \times R^3 \quad (4.12)
\]

where:

\[
KOHM = \text{resistivity reading from probe}
\]
CABLE = 0.2 K OHMS (from manual)
C(0) = 0.002
C(1) = 0.0013
C(2) = -0.0008
C(3) = 0.0003

Measurements were recorded every 25 meters from the top of the well casing to a
depth of about 5 m above the independently measured water level and then readings were
made every 1 m to the bottom of the well. The raw data were then entered into
spreadsheet files and converted to temperature values using Matlab. Appendix A
contains tables of depth, resistivity, and temperature for each of the wells that were
logged. The resulting temperature profiles were then plotted relative to mean sea level for
easy comparison (fig. 4.14). Epp and Halunen (1979) published temperature
measurements that were collected in 1974-76 in some of the same wells and these data
are plotted along with the most recent measurements in figures 4.15-4.17.

4.3.2. Discussion of Temperature Profiles

The profiles for GTW-3, MW-2, and Malama Ki Well clearly show a plume of
warmer, less-dense water floating on top of cooler, higher-density water. The Allison Well
profile looks like it is showing the same pattern but the well is not deep enough to provide
confirmation of this trend. GTW-3 is clearly the hottest shallow well, with temperatures
near the boiling point at the water table, and the hot water plume is also thickest (~ 40 m)
in this well. The MW-2 profile is similar in shape to the GTW-3 profile but cooler (61°C)
and thinner (~ 10 m). These profiles show an unstable temperature distribution which
Figure 4.14 Temperature profiles in East Rift Zone shallow wells, June 1, 1994.
Figure 4.15 Observed temperature profiles, GTW-3.
Figure 4.16 Observed temperature profiles, Kapoho Airstrip Well.
Figure 4.17 Observed temperature profiles, Malama Ki Well.
suggests that the wells are located down gradient and relatively close to a source(s) of hot water such as a fracture. The hot water may be buoyantly rising to the top of the water table and then is carried down gradient by the regional flow pattern. As the plume migrates away from the source, it should cool down and become thicker as the density difference with the shallow aquifer water becomes less pronounced. The profiles in the Malama Ki Well and the Allison Well suggest these wells are at a greater distance from a hot water source than GTW-3 and MW-2. The hottest water in Malama Ki Well is about 5 m below the water table. The plume was probably diluted by recharge from above along the flowpath between the hot water source and the well. Neither of the profiles from the Malama Ki and the Allison Wells is deep enough to show what happens to the base of the plume as the warm water migrates farther down gradient. The profiles from the Kapoho Airstrip Well and MW-1 are different because they do not show a clear temperature inversion. The MW-1 profile becomes steadily hotter with depth to the bottom of the profile. This profile seems to suggest that MW-1 is located directly above or slightly up gradient of a source of hot water or steam. The Kapoho Airstrip Well has a similar profile but the well is not deep enough to confirm any type of trend.

The plots comparing the most recent temperature profiles to those collected by Epp and Halunen in the mid-1970's show little variation over a 20 year time span. The most recent profiles fall mostly within the range of variations in the earlier profiles which were all collected within two years of each other. From these plots, it can be argued that the input to the shallow ground-water system in the vicinity of the 1955 eruptive center has remained at a steady-state condition for at least the last twenty years.
4.4. Chemical Characteristics of the Shallow System

Information about the ground-water geochemistry in the ERZ has been summarized in several recent reports including Scholl et al. (1993), Thomas and Janik (unpublished), Thomas (1994), and Novak (1995). Ground water can be geochemically divided into three regions; north flank waters, lower east rift zone waters, and south flank waters. In the north flank waters, dissolved solid concentrations are relatively low, having an average of 100 mg/kg and increasing to greater than 200 mg/kg near the coast. The dominant anion/cation pairs in north flank waters are sodium/calcium bicarbonate, but grade to sodium chloride near the coast (Swain, 1973). Tritium measurements from several north flank wells have been used to obtain aquifer residence-time estimates of less than 15 years and deuterium measurements indicate the water originally fell as rainfall at an altitude between 640 and 1,740 m (Scholl et al., 1993). Thomas (1994) analyzed samples collected from the Paradise Park Well (a north flank well) and considered the water to have chemistry typical of Hawaiian basal groundwater undisturbed by thermal influence. The general trend in the ground-water chemistry indicates that waters north of the rift contain extremely low dissolved solids concentrations and, as the rift zone is approached, the contribution of dissolved solids from naturally occurring discharge of geothermal fluids progressively increases (Thomas, 1994).

Water chemistry in the ERZ is more complicated because it appears to be the result of mixing of variable amounts of several distinct fluid types; freshwater, saltwater, thermally altered freshwater, and thermally altered saltwater. The wells from the northern
portion of the ERZ (Pahoa, MW-1, MW-3) can be grouped together based on sulfate concentrations. The wells penetrate waters which appear to be basal water heated by steam (Thomas, 1994). Wells from the southern portion of the rift (MW-2, GTW-3, Malama Ki Well, Allison Well, and Kapoho Airstrip Well) can be grouped together and appear to be comprised of freshwater and thermally altered saltwater. MW-2 and GTW-3 show the most intense alteration and are thought to be the closest to a source or sources of upflow into the shallow aquifer from a deeper geothermal reservoir. Several studies have shown that, in general, the ground water has immature compositions indicating that it has not had time to reach equilibrium with the aquifer rocks. Novak (1995) reported that it can be concluded that lower Puna wells and springs contain immature or slightly thermally altered waters whose residence times in the lower Puna ground water system have been relatively short. Three out of four tritium samples from the LERZ indicate an aquifer residence time of 10 to 20 years, and deuterium samples indicate recharge altitudes are between 400 to 850 m (Scholl et al., 1993).

The Kapoho Shaft Well, located in the tuff layer east of Kapoho Crater has a unique chemical signature. It is postulated that water in this well is derived from infiltration through Kapoho Crater into an ash bed which may locally perch the water. The water interacts with the ash which causes potassium, magnesium, and calcium to be leached out of the ash; ion balance is maintained by bicarbonate ion dissolved from the soil gas (Thomas, 1994). This local perched system probably has little connection to the larger rift zone flow system. Isotope studies of lake water indicate rather local and rapid recharge by cold water (Janik et al., 1994). As part of this study, simple temperature and conductivity profiles were collected at about 1.5-m depth intervals in the center of the lake.
to determine if a component of geothermal fluid is present (table 4.4). The profile shows a slightly higher temperature and conductivity at the lake surface indicating solar heating and evaporation to the atmosphere. There does not appear to be a measurable amount of geothermal fluid entering Green Lake.

Table 4.4. Green Lake temperature and conductivity profile, August 17, 1994

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Temperature (°C)</th>
<th>Conductivity (µmhos)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>28.3</td>
<td>8,669</td>
</tr>
<tr>
<td>1.5</td>
<td>28.4</td>
<td>9,158</td>
</tr>
<tr>
<td>3.0</td>
<td>28.1</td>
<td>8,547</td>
</tr>
<tr>
<td>4.6</td>
<td>27.3</td>
<td>5,617</td>
</tr>
<tr>
<td>4.9</td>
<td>26.8</td>
<td>3,541</td>
</tr>
</tbody>
</table>

Water chemistry in the south flank appears to be strongly influenced by the presence of thermally altered water that has moved down gradient from the rift zone. Two wells, Malama Ki Well and Allison Well, show chloride values of 5,400 and 2,040 mg/kg, respectively. Although Malama Ki Well is 3 km from the coast, it has a higher chloride concentration than Allison Well which is about 2 km from the coast. In a typical basal lens, chloride concentrations at the top of the water table usually decrease away from the coast where saltwater mixing is highest. This reversal in the trend of chloride concentration and the elevated silica concentrations and temperatures indicate that these wells are penetrating water which has a significant component of thermally altered water from the rift zone. Two other wells (Pulama and Keauohana) have lower chloride concentrations but elevated silica concentrations, which suggest that these wells receive a smaller amount of outflow from an up-gradient hydrothermal system (Novak, 1995). Puna shoreline warm springs have chemistry signatures much like the south flank wells.
although the water samples usually contain a high percentage of seawater due to their direct connection with the ocean.

Novak (1995) reports two basic processes which control the chemistry of lower ERZ and flank water; thermal saltwater mixing and rock dissolution. Thermal saltwater rises convectively through and mixes with the fresh overlying ground water, dominating the chemistry of high and medium chloride wells. Higher than ambient ion concentrations in the thermal waters suggest an increased temperature-dependent solubility of the major rock-forming constituents; silica, sodium, potassium, calcium, magnesium, and iron. The low chloride wells are dominated by rock dissolution. Thomas (1994) reported that temporal variations in ground-water chemistry are highly dependent on recharge events and that well chemistry within the rift may be more sensitive to recharge than outside the rift.

**Conceptual Model of the Shallow Flow System**

A high volume of recharge to dike-impounded aquifers in the UERZ and MERZ allow ground-water heads to build up as high as 200 m above sea level. The abundance of low-permeability intrusive bodies in the rift zone separate the flat-lying aa and pahoehoe lava flows into compartments, each containing freshwater to a depth of about 1 km below the ground surface where the porosity and permeability are very low and allow little or no ground-water flow. The porosity and permeability are most likely low due to compaction of the void space by the overlying rock, the presence of very dense intrusive bodies and the filling of void space by hydrothermal alteration. The presence of saltwater
at depth in the UERZ and MERZ is not likely but has not been confirmed or denied by sampling.

The rift zone appears to act as a flow barrier to ground water originating on the flanks of Mauna Loa north of the rift zone and diverts a large volume of ground water to submarine springs between Hilo and Cape Kumukahi. Isotope studies suggest that water in wells north of the rift originally entered the ground at altitudes higher than the summit of Kilauea, while water in wells in the LERZ originated from much lower altitudes. Moving down rift, the frequency of dike occurrences diminishes and ground water heads become lower. A portion of the flow probably travels perpendicular to the rift through higher-permeability fracture zones into the south flank lava flows and discharges at the coast. The fracture zones could possibly be transform fault-like features related to the stress on the south flank due to repeated magma injection along the rift zone, but their origin is still widely debated. The most prominent of these fracture zones is thought to lie between Puulena Crater and Puu Honuaula. Geophysical studies (Flanigan et al., 1986b, Kauahikaua, 1993) reveal the presence of low-resistivity areas south of the rift zone which have been suggested to be leaks of hydrothermal fluid migrating from the rift zone.

Hydrothermal fluid reaches the shallow aquifer through faults or fractures which extend from the ground surface to depths of as much as 2,100 m below sea level. At Puu Honuaula, there appears to be a set of fractures which define the boundaries of a graben (fig. 3.1). Periodic leveling surveys show that the area underlying the current geothermal production area has subsided as much a 0.6 m between 1958 and 1988. North of Puu Honuaula, a set of fractures up gradient of GTW-3 and possibly beneath MW-1 allow geothermal fluid to reach the shallow system. At MW-1 steam is heating the fresh ground
water which lowers its density and allows it to rise to the top of the water table. At GTW-3, with no evidence of steam heating, a different fracture must be discharging hot water with a component of thermally altered saltwater to the top of the water table where a plume spreads laterally down the rift. South of Puu Honuaula, on the southern edge of the graben, the same thing occurs but with the added complication of saltwater entering the system. Saltwater may be entering the deep geothermal system through a fracture zone where it is heated and buoyantly rises to the shallow system. Conversely, because this area is near the southern edge of the rift zone, the shallow aquifer may be permeable enough to allow saltwater to flow beneath the high-level impounded water. In this case, hot water from the fracture could be discharging directly into a transition zone and heating saltwater. The fractures are probably self-sealing due to deposition of hydrothermal alteration minerals and thus have a limited life-span. Little information exists about the age of fractures or the length of time they can continue to discharge before permeabilities become too low. The presence of thermally altered saltwater would tend to accelerate this process.

There are undoubtedly other comparable systems both up and down the rift based on temperatures measured in the Malama Ki, Allison, and Kapoho Airstrip Wells. Unfortunately, other than expensive well drilling, no reliable method of detecting these zones has been found to date. Similar conditions of faulting and recent magma intrusion occur at all of the 1955 and 1960-61 vent locations, but the lack of monitoring wells makes it difficult to confirm the existence of similar hydrothermal systems. The presence of steam vents at the bottom of Puulena Crater, and the high temperatures down gradient in the Malama Ki Well suggest hydrothermal action is still occurring in this area even though
magma has not reached the surface within 1.5 km of the crater for at least 400 years. The elevated concentration of sulfate in the Pahoa Wells suggest that there is an up-gradient source of steam heating the shallow ground water.

The less-frequent occurrence of intrusive dikes causes the permeability in the rift zone within 5-10 km of Cape Kumukahi to be about the same as the permeability outside the rift based on water-level and tidal-response measurements in wells. In this area, ground water is able to flow away from the rift towards the coast along the topographic gradient and discharge as basal, warm freshwater springs. Temperatures in most of the coastal springs are between 30-40°C.

The temperature and chemical concentration of ground water is highly dependent on recharge and daily tidal fluctuations of water in the aquifer. After heavy rainfall events, due to dilution, observed ground-water temperatures and ion concentrations decrease, and during dry periods observed values increase. The lag time between rainfall and well response depends on the lithology of the unsaturated zone in the vicinity of the well. North of the rift zone, the Paradise Park Well responded within about 10 hours to a heavy recharge event. South of the rift, the presence of a low-permeability tuff layer(s) greatly reduces the infiltration rate of the recharge and the recharge-response relationship is much more complicated. Daily fluctuations of water level in the aquifer are caused by the propagation of pressure pulses from ocean tides through the aquifer matrix. Higher fluctuations are found nearer the coast and the rift zone is an effective barrier to tidal propagation perpendicularly to the plane of the intrusive dikes.
5. COUPLED FLOW, ENERGY, AND SOLUTE TRANSPORT MODELING

This chapter explains the concept of the numerical models which solve the basic mathematical equations of fluid flow, energy transport, and solute transport. The numerical models are verified by comparing the results of simple simulations to problems with known analytical solutions. Further verifications are provided by comparing the results of two different models used to solve an identical problem.

5.1. Mathematical Formulation

In developing a ground-water-flow or solute-transport model, the analyst first translates the physical and chemical processes governing the behavior of a system into mathematical terms (Istok, 1989). The mathematical model is, in general, a collection of partial differential equations and auxiliary conditions which can be solved analytically or numerically. The mathematical equations governing the time-dependent transport of solutes and energy in ground water are sufficiently complex that numerical models are usually necessary to solve them. The following sections briefly describe the mathematical equations used in this study and two different numerical, computer models incorporating these equations which have been used to simulate the flow and transport characteristics of ground water in the Kilauea ERZ.
5.1.1. Flow Equation

Ground-water flow is usually simulated through solution of a fluid mass balance equation for a non-deformed solid matrix:

$$\frac{\partial (\varepsilon S_w \rho_f)}{\partial t} = - \nabla \cdot (\varepsilon S_w \rho_f \mathbf{v}) + Q$$  \hspace{1cm} (5.1)

where:

- $\varepsilon$ = matrix porosity [dim]
- $S_w$ = water saturation [dim]
- $\rho_f$ = fluid density [M/L^3]
- $\mathbf{v}$ = average interstitial flow velocity [L/T]
- $Q$ = fluid mass source [M/L^3·T]

The term on the left of (5.1) represents the total time change in fluid mass contained in the void space of a porous medium. The first term on the right represents the change in fluid mass due to the excess of fluid inflows over outflows in the volume. The $Q$ term represents any external addition or subtraction of fluid mass such as from an injection or pumping well. For this study, only the fully saturated zone is considered so $S_w$ is always equal to 1.0. For describing ground-water flow in the KERZ, which contains freshwater/saltwater mixtures and variable-temperature fluids, neither the porosity nor the fluid density are constant in time or space. The porosity of the aquifer is dependent on pressure, and fluid density is dependent on pressure, solute concentration, and temperature. The time derivative term of equation (5.1) can be expanded to show the dependence of density and porosity on pressure, solute concentration, and temperature.
as follows:

\[
\frac{\partial (\varepsilon \rho)}{\partial t} = \rho S_{op} \frac{\partial (\varepsilon \rho)}{\partial t} + \varepsilon \frac{\partial \rho}{\partial C} \frac{\partial C}{\partial t} + \varepsilon \frac{\partial \rho}{\partial T} \frac{\partial T}{\partial t} \tag{5.2}
\]

The first term on the right of (5.2) represents the dependence of the fluid density and porosity on changes in pressure. Storativity is defined as follows (Bear, 1979)

\[
S_{op} \equiv \frac{1}{\rho} \frac{\partial (\varepsilon \rho)}{\partial p} = (1 - \varepsilon) \alpha + \varepsilon \beta \tag{5.3}
\]

where:

\[
\alpha = \text{porous matrix compressibility} \quad [\text{M/L}^2\cdot\text{T}^{-1}]
\]
\[
\beta = \text{fluid compressibility} \quad [\text{M/L}^2\cdot\text{T}^{-1}]
\]

The second term on the right of (5.2) represents the dependence of density on the fluid dissolved ion concentration and the third term represents the dependence of density on the fluid temperature.

The flow velocity term in (5.1) is commonly expressed using Darcy’s Law for fluid flow in porous media:

\[
\mathbf{v} = - \left( \frac{k}{\varepsilon \mu} \right) \cdot (\nabla p - \rho g) \tag{5.4}
\]

where:

\[
k = \text{solid matrix permeability} \quad [\text{L}^2]
\]
\[
g = \text{gravitational acceleration vector} \quad [\text{L/T}^2]
\]
\[
\mu = \text{fluid viscosity} \quad [\text{M/L}\cdot\text{T}]
\]
\[
p = \text{fluid pressure} \quad [\text{M/L}\cdot\text{T}^2].
\]
Substituting (5.2), (5.3), and (5.4) into (5.1) yields the partial differential equation for density dependent flow:

\[
\rho S_{op} \frac{\partial p}{\partial t} + \varepsilon \frac{\partial \rho}{\partial t} \frac{\partial C}{\partial t} + \varepsilon \frac{\partial T}{\partial t} \frac{\partial C}{\partial t} = \nabla \cdot \left\{ \frac{k \rho}{\mu} \cdot (\nabla p - \rho g) \right\} = q \tag{5.5}
\]

Bear (1972) list six assumptions that the development of (5.5) are based on;

1) the solid matrix is homogeneous, nondeformable and chemically inert with respect to the fluid;

2) the fluid is single phase and Newtonian;

3) the fluid is a binary system with a molecular diffusion coefficient \( D_d \);

4) the flow is in the laminar range;

5) no chemical reactions take place among the fluid’s species; and

6) no heat sources or sinks exist in the fluid.

### 5.1.2. Solute Transport Equation

The mass balance equation for a non-sorbing, single species dissolved in solution is expressed as follows:

\[
\frac{\partial (\varepsilon \rho C)}{\partial t} = - \nabla \cdot (\varepsilon \rho \nu C) + \nabla \cdot (\varepsilon \rho (D_m I + D) \cdot \nabla C) + QC^* \tag{5.6}
\]

where:
The term on the left of (5.6) represents the total changes of solute mass over time in a volume due to the mechanisms represented by the terms on the right. The first term on the right represents the advection of solute mass along with the fluid at the average Darcy velocity (advective flux). The second term represents the contribution of solute diffusion and dispersion to the local changes in solute mass (hydrodynamic dispersive flux). The diffusion term represents a physical process which can often be ignored at field scale. The dispersion term represents an approximation of the effect of solute advective mixing in irregular flow fields (tortuous pathways) which is not accounted for in the advection term. It is treated in the same manner as diffusion (as a Fickian process) for purely practical reasons because on the macroscopic scale, the outcome is the same for both processes. Logically, mechanical dispersion should be reflected in the velocity term, but it is not possible to do this in a simple way (Domenico and Schwartz, 1990). The final term represents solute mass added or removed through a fluid source or sink. The assumptions listed for the development of (5.5) apply to (5.6) as well.

Two of the terms in (5.6) contain the flow velocity (v), which must be determined by first solving the fluid flow equation. But the fluid flow equation contains density and viscosity terms which are dependent on the solute concentration. Therefore, these two equations must be solved simultaneously in an iterative fashion until an acceptable convergence limit is obtained.
The most troublesome part of solving (5.6) arises from uncertainties in the dispersion term. Two of the models commonly used to describe dispersion depend on whether the aquifer has isotropic or anisotropic permeability. The anisotropic-media dispersion model is represented by the two-dimensional dispersion tensor:

\[
D = \begin{bmatrix}
D_{xx} & D_{xy} \\
D_{yx} & D_{yy}
\end{bmatrix}
\]  

(5.7)

where:

\[
D_{xx} = \left(\frac{1}{v^2}\right)(d_L v_x^2 + d_T v_y^2)
\]  

(5.8)

\[
D_{yy} = \left(\frac{1}{v^2}\right)(d_T v_x^2 + d_L v_y^2)
\]  

(5.9)

\[
D_{xy} = D_{yx} = \left(\frac{1}{v^2}\right)(d_L - d_T)(v_x v_y)
\]  

(5.10)

\[
d_L = \alpha_L v
\]  

(5.11)

\[
d_T = \alpha_T v
\]  

(5.12)

and:

\[
\alpha_L = \text{longitudinal dispersivity length} \quad [L]
\]

\[
\alpha_T = \text{transverse dispersivity length} \quad [L]
\]

The anisotropic-dispersion model (Voss, 1984) assumes two principal directions of longitudinal dispersivity (\(\alpha_{L\text{max}}\) and \(\alpha_{L\text{min}}\)) aligned with principal directions of permeability.
and $\alpha_L$ is determined from an equation for an ellipse:

$$\frac{1}{\alpha_L} = \frac{\cos^2 \theta_{kv}}{\alpha_{Lmax}} + \frac{\sin^2 \theta_{kv}}{\alpha_{Lmin}}$$

(5.13)

where:

$$\theta_{kv} = \text{angle of flow with respect to } \alpha_{Lmax}.$$ 

Transverse dispersivity would also tend to be dependent on the flow direction, but because it typically is only a small fraction of longitudinal dispersivity, especially in anisotropic media (Gelhar and Axness, 1983), its variability is ignored here. Recent studies have suggested more complex ways of handling the scale-dependent nature of dispersion (i.e. using fractals and stochastic approaches; see Neuman, 1990), but current numerical models have not yet been integrated with these findings.

### 5.1.3. Energy Transport Equation

The equation for the mass balance of changes in stored energy with various energy fluxes (modified from eq. 10.7.23 in Bear, 1972) is expressed as follows:

$$\rho_e c_e \frac{\partial T}{\partial t} = -\rho c_f \nabla \cdot T + \nabla \cdot \left[ \lambda_e I \cdot \nabla T \right] -$$

$$\rho c_f \nabla \cdot (\varepsilon E \cdot \nabla T) + Q c_f T^* + \rho_s H_i$$

(5.14)

where:

$$\rho_s = \text{effective density of saturated porous media} \quad [M/L^3]$$
\( c_e = \) effective specific heat capacity of saturated porous media
\( T = \) temperature of fluid
\( c_f = \) specific heat capacity of fluid
\( v = \) average uniform flow velocity
\( \lambda_e = \) effective thermal conductivity of saturated porous media
\( I = \) identity tensor (ones on diagonal, zeroes elsewhere)
\( E = \) dispersion tensor
\( Q = \) fluid mass source
\( T^* = \) temperature of an external source fluid
\( \rho_s = \) density of solid matrix
\( H_i = \) source of energy without fluid

The term on the left of (5.14) represents the total change over time in energy stored in both the solid matrix and fluid per unit total volume. The first term on the right represents the changes in locally stored energy through the advection of energy by the fluid at the average uniform flow velocity (advective or convective flux). The second term represents changes in locally stored energy due to conduction, described by Fourier’s law, in both the fluid and solid matrix (conductive flux). This term is analogous to molecular diffusivity in solute transport. The third term on the right-hand side of (5.14) represents an approximation of the contributions of irregular flows and mixing not accounted for in the advective term (dispersive flux). Although the dispersive flux is essentially an advective process, its effects are macroscopically equivalent to conductive transport. Therefore it is handled using the form of Fourier’s law. The fourth terms represents changes in stored energy due to sources of fluid at known temperatures, and the fifth term represents changes in stored energy due to sources of external energy without fluid. The assumptions listed for (5.5) apply to (5.14) as well.
In the development of (5.14) the temperature difference between the fluid and the solid matrix is neglected. This assumption has been shown to be acceptable for modeling hydrothermal systems because of the rapid equilibration between the fluid and solid matrix. For a fractured rock where the solid blocks have a thickness of 0.1 m, the estimated time required for the temperature of the solid to approach the temperature of the fluid is about two hours (Houpert et al., 1965). Basalt aquifers, in general, would have "block sizes" less than 0.1 m. Sorey (1993) estimates particle velocities in the ERZ might be, at most, about 3-30 m/d. Therefore, in two hours, a packet of water should only travel between 0.25-2.5 m, providing ample time for equilibration with the surrounding matrix.

The energy mass balance equation contains velocity in both the advective and dispersive terms so it must be solved simultaneously with the fluid mass balance and solute mass balance equations in an iterative fashion.

The effective density and heat capacity of the saturated media is determined from the following (Bear, 1972):

$$\rho_e c_e = \epsilon \rho_f c_f + (1 - \epsilon) \rho_s c_s$$

(5.15)

where:

$$c_s = \text{specific heat capacity of solid matrix} \quad [\text{E/M}^\circ\text{C}]$$

and the other terms are as described previously.

The effective thermal conductivity is usually represented by a simple parallel conduction model in which conduction occurs separately but simultaneously in both the
solid and the fluid (Bear, 1972):

\[ \lambda_e = \varepsilon \lambda_f + (1 - \varepsilon) \lambda_s \]  \hspace{1cm} (5.16)

where:

\[ \lambda_f = \text{thermal conductivity of fluid} \quad [\text{E/T} \cdot \text{L} \cdot ^\circ \text{C}] \]

\[ \lambda_s = \text{thermal conductivity of solid matrix} \quad [\text{E/T} \cdot \text{L} \cdot ^\circ \text{C}] \]

Thermal conductivity is assumed to be isotropic in the porous media (\( \lambda_{ex} = \lambda_{ez} \)).

The energy dispersion tensor for energy transport is analogous to the solute dispersion term (5.7) described previously with one basic difference. The concept of tortuosity applies only in solute transport because the transfer takes place through the void space only. Energy transfer occurs in both the fluid and solid portions of the porous matrix. But based on experimental data, the passage of heat through the solid phase is insignificant when compared to advective heat transfer (Bear, 1972), so both solute and energy dispersion can be described through essentially the same process.

### 5.2. CFEST Numerical Model

CFEST is a finite-element code for three-dimensional analysis of hydrologic flow, heat transport, and single-constituent solute transport in subsurface confined environments at either the regional or local scale (Cole, 1988). The version of the code used for this study was originally produced for use on a CRAY X-MP supercomputer. Initially, the San Diego Supercomputing facility was to be utilized for this study, however,
turnaround time for model results was unacceptably slow so this idea was abandoned. The Unix workstations at SOEST provided equivalent turnaround time, hence SOEST SPARC 10 workstations were used. The CFEST package contained many programs, subroutines, and Unix shells for pre-processing and post-processing of input and output files for the code. Most of the auxiliary programs had little usefulness over the course of the study. The main program ran fairly slowly, especially when both energy and solute transport options were used. Several major bugs in the code were found and fixed, some with the help of one of the code authors (Charles R. Cole, Battelle PNL) but most without. The code had a major drawback in that it required huge amounts of computer storage space to save all of the output files produced by a long simulation. The program saved output from every time step for restarting capabilities and there was no way to turn this feature off. The following section will describe some of the features and methods used in the code and some modifications made to the code. For a full description of the code and features, the reader is referred to the CFEST manual (Cole, 1988).

5.2.1. Features and Capabilities

The model was originally written for confined aquifers, but it handles water-table simulations with an external subroutine which updates the thickness of the top element of the finite-element mesh to match the elevation of the water table as it changes over time. The user's manual recommends using this subroutine whenever the approximate head changes in the top row of elements exceed five percent of the aquifer thickness. CFEST will allow the user to simulate the flow of fluid with both solute and energy transport. For
solute transport, it allows for concentration-dependent fluid density and viscosity, linear adsorption and decay of solutes, solute sources and sinks, and salt dissolution in the aquifer. For energy transport it includes temperature-dependent fluid density and viscosity, sources of heat from injection wells, and sources of heat but no fluid from overlying and underlying confining beds.

Fluid density is described as a function of solute concentration, temperature, and pressure using the following equation of state:

$$\rho(p,C,T) = a_1(p-p_0) + a_3C + \rho(T)$$  \hspace{1cm} (5.18)

The first term on the right describes the changes of density due to changes in the pressure exerted on the fluid. This term is the least significant of the three factors controlling fluid density. The components of this term are:

- $a_1 = \rho_0 C_w = \text{base density} \times \text{fluid compressibility}$  \hspace{1cm} [M/L$^2$]
- $\rho_0 = \text{base density of fluid at } p_0, T_0 \text{ and } C = 0$  \hspace{1cm} [M/L$^3$]
- $p = \text{local pressure on fluid}$  \hspace{1cm} [M/L.$T^2$]
- $p_0 = \text{base pressure on fluid at } T_0 \text{ and } C = 0$  \hspace{1cm} [M/L.$T^2$]

The second term on the right describes the changes in fluid density due to the concentration of solutes in the fluid where:

- $C = \text{local concentration of fluid}$  \hspace{1cm} [M/L$^3$]
- $a_3 = \rho_0 C_o = \text{base density} \times \text{fluid composition density ratio}$  \hspace{1cm} [dim]
and $C_c$ is determined from:

$$C_c = \frac{\rho_1(C_1, T_0, p_0) - \rho_o}{\rho_o C_1}$$  \hspace{1cm} (5.19)$$

where:

$\rho_1$ = density of fluid at a known concentration $C_1$,

temperature $T_0$ and base pressure $p_o$ \text{[M/L$^3$]}

For a simulation of saltwater-freshwater interaction:

$$C_c = \frac{1022.5 \frac{kg}{m^3} - 996.9 \frac{kg}{m^3}}{996.9 \frac{kg}{m^3} \times 18900 \frac{mg}{l}} = 1.36 \times 10^{-6} \left[ \frac{mg}{l} \right]^{-1}$$  \hspace{1cm} (5.20)$$

The third term of (5.18) represents the dependence of density on temperature and is represented in CFEST with the following two relationships:

$25^\circ C \leq T \leq 100^\circ C$:

$$\rho(T) = 996.9 \times \left\{ 1 - 3.17 \times 10^{-4}(T - 25) - 2.56 \times 10^{-6}(T - 25)^2 \right\}$$  \hspace{1cm} (5.21)$$

$T \leq 25^\circ C$:

$$\rho(T) = 996.9 \{1 - 1.87 \times 10^{-4}(T - 25)\}$$  \hspace{1cm} (5.22)$$

Equations 5.21 and 5.22 are formulated in units of kilograms, meters and degrees Celsius.

Variable fluid viscosity can be modeled by three different options in the CFEST code: 1) as a function of temperature only using an exponential model or tabulated...
temperature-viscosity data, 2) as a function of temperature and concentration using tabulated concentration-viscosity data and an exponential temperature-viscosity model, or 3) as a function of concentration only using tabulated concentration-viscosity data. The first option was chosen for this study with tabulated temperature-viscosity data for freshwater taken from the literature. The effects of concentration on viscosity were ignored which leads to about a seven percent error in the estimate of viscosity between the freshwater and saltwater endmembers of the fluid represented in the simulations.

CFEST also has the option for choosing constant or temperature-dependent values for fluid specific heat and fluid thermal expansion. The temperature-dependent option has a tabular look-up procedure for fluid temperatures up to 300°C. The effect of using variable rather than constant thermal expansivity is probably minor for the relatively low temperatures simulated in this study but it is more complete. For a simple convection-cell model, the effect of using the variable option would be one of increasing the instability of a water body heated from below by reducing the critical Rayleigh number for the onset of convection in a cell (Strauss and Schubert, 1977). For all simulations, unless otherwise stated, the temperature-dependent option was used for a more realistic simulation.

5.2.2. Code Structure

CFEST is divided into 5 main programs for solving the governing equation of flow and transport. The first program (LPROG1) reads the main input data and estimates the number of unknowns in the problem. The second program (LPROG2) estimates the
Jacobians and the derivatives at each quadrature point and surface on an element. The third program (LBAND) calculates the element locations for the formulation of a system matrix to be solved by the equation solver. The fourth program (LPROG3I) reads time steps, recharge rates and pumping stresses for the simulation. The fifth and main program (LPROG3) calculates pressure, temperature, concentration and density for the simulation. Figure 5.1 shows the code flow chart.

The code first solves for pressure using the initial conditions (or a previous time step) of temperature and/or concentration. The equation for temperature is then solved using the new pressure solution and the previous or initial concentration distribution. Finally, the concentration is determined from the new pressure and temperature distributions. The pressure for the next time step is solved from the new temperature and concentration solutions and so on. For each temperature and concentration solution, density convergence is tested as follows:

\[
\frac{\Delta \rho^T + \Delta \rho^C}{\rho_0} \leq 0.001
\]

where:
- \(\Delta \rho^T\) = the maximum change in density at any node because of change in temperature over an iteration
- \(\Delta \rho^C\) = the maximum change in density at any node because of change in concentration over an iteration.

The iterations continue until the tolerance level in (5.23) is met.
LPROG1: Reads the Geometric and Physical Parameters and Generates the Binary Files for Other Programs. LPROG1 Requires an Input File.

LPROG2: Generates Constant Integration Parameters for Element Matrix Formation. LPROG2 is Executed Only Once for a Given Problem; No Input File is Required.


LPROG3: Main Program for Cyclic Estimation of Head, Temperature, and Concentration (Figure 5-9). Through Control File the User Selects Various Combinations of Parameters for Simulation, Inclusion of Heat Transfer to Confining Rocks and Subdivision of Transport Time Steps for a Given Flow Timestep.

Figure 5.1 Flowchart of CFEST model. (from Cole et al., 1988)
Several modifications were made to the original code. Some modifications were to fix some bugs that were found and some were to make the code run better. The biggest problem found was that the original code did not restart from the middle of a temperature simulation correctly. It did not externally save the three vectors that contained the time derivative terms of p, T, and C changes from the last time step completed. Therefore, when a simulation was restarted these terms were not available and the resulting simulations were incorrect. A simulation of hot water injection would incorrectly start to cool down upon restarting rather than continue to heat up. To correct this problem, it was necessary to add a few lines of code to save this information externally and read it in for a restarted simulation.

Another problem with the code was that it required huge amounts of storage space to save the results of each time step in a simulation. For simulations involving a freshwater/saltwater transition zone, fairly small time steps are needed to get correct results without considerable numerical dispersion. To illustrate this problem, consider a simulation where hot water has been injected into a freshwater/saltwater transition zone for 40 years. This simulation requires about 40,000 time steps at 8.3 hours/timestep. A CFEST run of just 600 timesteps requires over 60 MB of disk storage. So it is readily obvious that it is impossible to complete an entire simulation without having restart capabilities. It was necessary to modify the code to take the final timestep of a simulation and return it into the storage files as the first timestep to "trick" the code into continuing as if it never stopped.
An additional problem with the code was the way it handled storativity for an unconfined-aquifer simulation. If the aquifer and fluid compressibilities are entered, CFEST calculates the specific storage for each element from (5.3). Otherwise, it calculates the compressibilities based on the entered specific storage. This method works fine for all elements except the top layer where the specific storage value needs to be treated as a specific yield with a value on the order of five percent. For these elements, CFEST calculated anomalously high values of compressibility which caused incorrect solutions in the top row of elements. A simple modification to the code (subroutine LKXYZ) made it possible to correctly represent the storage in the top row of elements.

Another problem with using the code for water-table simulations appeared in the program for updating the thickness of the top element of the mesh (LHW KurtZ). This subroutine was written for constant-density simulations where the dependent variable was head instead of pressure. Therefore, the code had to be modified to convert the pressure in the top element to an equivalent head before updating the new mesh coordinates.

And finally, a modification was made to the subroutine which calculated the dispersive flux (LQDERVI). The code originally determined the dispersion tensor from the isotropic-media dispersion model represented by (5.8-5.13). A flow-dependent dispersion model (eq. 5.14) based on the one used in the SUTRA code (see SUTRA Numerical Model section) was added to the CFEST code. This approach is thought to be more appropriate for modeling island aquifer transition zones (Souza and Voss, 1987).
5.3. SUTRA Numerical Model

SUTRA (Voss, 1984) is a finite-element saturated-unsaturated, fluid density-dependent ground-water flow model with either energy or chemically reactive single species solute transport. SUTRA can be used for either two-dimensional areal or cross-sectional modeling. The code is written in Fortran and can be run on either pc's or Unix workstations. For this study, all simulations were done on the SOEST computer network. The code is straightforward and very easy to use and has been well documented in studies of island aquifer modeling (Voss and Souza, 1986, Griggs, 1989, Underwood, 1990).

5.3.1. Features and Capabilities

The model is modular in design and can be easily modified to add enhancements to the solving routines. For solute transport it allows for concentration-dependent fluid density, several options for equilibrium sorption, first-order or zero-order decay of solutes and solute sources. For energy transport it includes temperature-dependent fluid density and viscosity capabilities, and allows sources of heat from injection wells. Fluid density is described as an approximate function of solute concentration or temperature using one of the following equations of state.
for energy transport:

$$\rho(T) = \rho_o + \frac{\partial \rho}{\partial T}(T - T_o)$$  \hspace{1cm} (5.24)

where:

- $\rho_o$ = base fluid density at $T = T_o$  \hspace{1cm} [M/L^3]
- $T_o$ = base fluid temperature \hspace{1cm} [°C]
- $\partial \rho/\partial T$ = constant value of density change with temperature, \hspace{1cm} [M/L^3°C]

or for solute transport:

$$\rho(C) = \rho_o + \frac{\partial \rho}{\partial C}(C - C_o)$$  \hspace{1cm} (5.25)

where:

- $\rho_o$ = base fluid density at $C = C_o$ \hspace{1cm} [M/L^3]
- $C_o$ = base fluid solute concentration \hspace{1cm} [M/M]
- $\partial \rho/\partial C$ = constant value of density change with concentration \hspace{1cm} [M/L^3]

Fluid viscosity is dependent on temperature through the following equation of state for energy transport:

$$\mu(T) = (239.4 \times 10^{-7}) \times 10^{(\frac{248.57}{T+133.75})}$$  \hspace{1cm} (5.26)

with the units being entered in [kg/m·s]. For solute transport, viscosity is taken to be constant.
5.3.2. Code Structure

The SUTRA code is contained in one program with many subroutines for reading and writing data, updating time dependent parameters and solving the governing equations. Figure 5.2 shows a flow chart of the functions and routines for the code.

5.3.3. Code Modifications

Only minor modifications were made to the SUTRA code because no bugs or problems were encountered. In order to simulate transient conditions in an aquifer, modifications were added to the code to allow time-varying recharge to be read from an external file.
Figure 5.2 Flowchart of SUTRA model. (from Underwood, 1990)
5.4. Numerical Model Verification

5.4.1. Solute Transport

In order to be certain that a numerical model is correctly solving the flow and transport equations, the results of model simulations must be compared to results predicted by analytical solutions. Both CFEST and SUTRA user's manuals include several examples of verification for flow and transport problems. CFEST fluid-flow solutions have been verified with four different simple analytical solutions and the results are acceptable. The solute-transport simulation capabilities are also verified against two simple analytical solutions for constant density fluid. The SUTRA documentation presents similar verifications including one for fluid flow and two for solute transport, one of which is for a variable-density fluid. All of the verifications are acceptable. Additional verifications of the CFEST code were performed here to match solute-transport simulations to analytical solutions, including one for a variable density fluid.

The Ogata-Banks equation is a solution to the one-dimensional dispersion-advection equation (Domenico and Schwartz, 1990):

\[
C(x, t) = \left( \frac{C_0}{2} \right) \left\{ \text{erfc} \left( \frac{x - vt}{2\sqrt{Dt}} \right) + \exp \left( \frac{vx}{D} \right) \text{erfc} \left( \frac{x + vt}{2\sqrt{Dt}} \right) \right\} \tag{5.26}
\]
where:

\[ \begin{align*}
C_0 &= \text{initial concentration of solute in fluid} \quad [M/L^3] \\
\chi &= \text{distance from source} \quad [L] \\
v &= \text{Darcy velocity of fluid} \quad [L/T] \\
t &= \text{time} \quad [T] \\
D &= \text{dispersion (} \alpha_x v \text{), ignoring diffusion} \quad [L^2/T] \\
\alpha_x &= \text{dispersivity length} \quad [L]
\end{align*} \]

This solution is for the conditions:

\[ C(0,t) = C_0 \quad \text{and} \quad C(x,0) = 0. \]

A simple 20 x 2 element mesh was created to solve this problem numerically using CFEST. The vertical element spacing was about 4.1 m. For the simulation, the base of the two-element-wide column was set at a constant head of 1.1 m and a constant chloride concentration of 100 mg/kg. The top nodes of the mesh were held at a specified head of 1.0 m. Longitudinal dispersivity length was set at a relatively high value of 60 m to insure that the numerical model could handle an extreme case. Viscosity and density were held constant. The observation point was 97 m (x) from the source. An actual flow velocity was obtained from the CFEST flow velocity subroutine and multiplied by the porosity (0.05) to get a Darcy velocity (1.12 \times 10^{-7} \text{ m/s}) to be used in the analytical solution. Figure 5.3 shows a very good comparison between the exact analytical solution and the modeled solution for a two-year simulation.

For another model verification, consider steady-state freshwater flow in the isotropic unconfined aquifer in figure 5.4 which penetrates into the ocean. Using Dupuit's
Figure 5.3 Ogata-Banks analytical solution and CFEST simulation.
Figure 5.4: Shape of freshwater lens based on Ghyben-Herzberg approximation.
assumption of horizontal flow (and vertical equipotentials), continuity leads to:

\[ Qx = - K \frac{dh}{dx} (h + z) \]  \hspace{1cm} (5.27)

By the Ghyben-Herzberg (isostatic) relationship:

\[ z = \delta h; \quad \delta = \frac{\rho_f}{(\rho_s - \rho_f)}; \quad z + h = \delta h + h = (1 + \delta)h \]  \hspace{1cm} (5.28)

substituting (5.28) into (5.27) yields:

\[ Qx = - K \frac{dh}{dx} (1 + \delta)h \]  \hspace{1cm} (5.29)

Rearranging terms in (5.29) and integrating from \( x=x \) to \( x=L \) and \( h=h \) to \( h=0 \):

\[ \frac{-Q}{(1+\delta)K} \int_{x=x}^{x=L} x \, dx = \int_{h=\delta}^{0} h \, dh \]  \hspace{1cm} (5.30)

and:

\[ \frac{-Q}{(1+\delta)K} \left[ \frac{1}{2} x^2 \right]_{x}^{L} = \left[ \frac{1}{2} h^2 \right]_{\delta}^{0} \]  \hspace{1cm} (5.31)

leads to:

\[ \frac{-Q}{(1+\delta)K} (x^2 - L^2) = \left( \frac{z^2}{\delta^2} \right) \]  \hspace{1cm} (5.32)
and:

\[ z^2 = \frac{Q\delta^2}{(1 + \delta)K} (L^2 - x^2) \]  

(5.33)

For an anisotropic aquifer, substitute \( K^* \) for \( K \) where:

\[ K^* = K_x \sqrt{\frac{K_x}{K_z}} \]  

(5.34)

A 2,684-element cross-sectional mesh oriented perpendicular to the coastline was used to numerically model this situation. The details of the mesh and the aquifer characteristics will be discussed later in the regional modeling section and the parameters used are listed in table 6.1. For the first case, the analytical equation was solved assuming an isotropic aquifer and either freshwater or saltwater in the aquifer. The horizontal and vertical permeability for freshwater was equivalent to 100 m/d and for saltwater was 95.8 m/d. Because an actual basal-lens system contains water with a continuous range of freshwater to saltwater density and viscosity, these two endmembers should bracket the actual case. From eq. (5.34), it is readily apparent that the depth of the interface (\( z \)) is inversely proportional to the square root of hydraulic conductivity (\( K \)), and lowering \( K \) by 4.2 percent will increase the depth of the interface by 2.05 percent. When the equation is solved, the saltwater endmember produces a lens that is about 2 percent deeper than the freshwater case (fig. 5.5).

The CFEST simulation of this problem produced a reasonably good match of the 50-percent seawater chloride value to the predicted position of the interface based on the
\[ z^2 = \frac{Q\delta^2}{K(1+\delta)}(L^2 - x^2) \]

K (fresh) = 100.0 m/d ....
K(salt) = 95.8 m/d ....
Q = .0017 m/d ....
\( \delta = 40 \)
L = 9500 m

Figure 5.5 Analytical solution and simulated solution, isotropic aquifer.
analytical solution. The simulated position is slightly lower in the interior of the lens and slightly higher near the coast. The differences can be attributed to the effects of numerical dispersion which are due to coarser element spacing at depth in the interior of the lens and increased flow velocities at the coast. Overall, the model does an acceptable job of simulating a basal lens system. For the second case, the analytical solution was solved for an aquifer with an anisotropy ratio of 2:1. When (5.35) is used to determine K, it has the effect of making the interface depth inversely proportional to the fourth root of the anisotropy ratio. For a ratio of 2:1, the depth of the interface will increase by $\sqrt[4]{2}$ or 1.19 (fig. 5.6). The CFEST simulation again produces a reasonable match for the anisotropic case.

5.4.2. Energy Transport

For CFEST, five different energy-transport verifications are presented, all of which are for constant density and viscosity scenarios. These verifications are acceptable but they are for fairly simple steady-state problems with constant values of density and viscosity. The SUTRA documentation provides one verification for transient transport of energy with an approximate analytical solution. The match is fairly good and the SUTRA result is stated to be possibly even more accurate than the approximate analytical solution because of the fine discretization employed in the SUTRA mesh (Voss, 1984). According to Cliff Voss (oral commun., 1995) early CFEST simulations involving energy-transport solutions with variable-density fluid initially were wrong and the code was corrected so that CFEST results would match results of SUTRA simulations. With this problem in mind, it
\[ z^2 = \frac{Q^6^2}{K^*(1 + \delta)} (L^2 - x^2) \]

- \( K_x \) (fresh) = 100.0 m/d, \( K_z = 50 \) m/d
- \( K_x \) (salt) = 95.8 m/d, \( K_z = 47 \) m/d
- \( Q = 0.0017 \) m/d
- \( \delta = 40 \)
- \( L = 9500 \) m
- \( K^* = ((K_x/K_z)^{0.5}) \times K_z \)

Figure 5.6 Analytical solution and CFEST simulation, anisotropic aquifer.
seemed obvious that additional verifications of CFEST were necessary before it could be completely trusted.

Turcotte and Schubert (1982, p. 401, eq. 9-113) present a one-dimensional, steady-state analytical solution for temperature as a function of depth of upwelling fluid above an intrusive body:

\[
T = T_r - (T_r - T_0) \exp \left( \frac{\rho f c_f f_0}{\lambda_m} y \right) \tag{5.35}
\]

where:

\[
\begin{align*}
T_r &= \text{reservoir (base) temperature} \quad [\degree C] \\
T_0 &= \text{temperature at surface} \quad [\degree C] \\
\rho_f &= \text{fluid density (constant)} \quad [\text{M/}^3] \\
c_{pf} &= \text{fluid specific heat} \quad [\text{E/M}.\degree \text{C}] \\
\nu &= \text{Darcy velocity (constant)} \quad [\text{L/T}] \\
\lambda_m &= \text{solid matrix thermal conductivity} \quad [\text{E/T-L}.\degree \text{C}] \\
y &= \text{depth} \quad [\text{L}]
\end{align*}
\]

A simple 20 x 2 element mesh was created to solve this problem numerically using CFEST. The vertical element spacing was about 4.1 m and extended to a depth of 650 m. For the simulation, the base of the two-element-wide column was set at a constant temperature of 150°C and the top nodes of the mesh were held at a constant temperature of 24°C. The physical constants were set as follows: \( \rho_f = 996.9 \text{ kg/m}^3 \), \( c_{pf} = 4182 \text{ J/kg}.\degree \text{C} \), \( \lambda_m = 2.0 \text{ J/m}.\degree \text{C} \). A constant vertical flow gradient was achieved by setting specified pressures at the top and bottom of the mesh, and an actual flow velocity was obtained from the CFEST flow-velocity subroutine and multiplied by the porosity (0.05) to get a Darcy velocity \( 1.813 \times 10^{-8} \text{ m/s} \) to be used in the analytical solution. Figure 5.7 shows very good agreement between the exact analytical solution and the CFEST solution.
Figure 5.7 One-dimensional advection solution and CFEST simulation.
A group of transient, energy-transport simulations were performed to compare the results between SUTRA and CFEST using the various simulation options available with the codes. The problem involved the simulation of vertical transport of energy in an upwelling fluid. Identical 2 x 20-element meshes were created for each model and the same physical properties and time-step lengths were used so each numerical code would be solving the identical problem. Each element was 4.21 m high and 5 m wide, the mesh represented an aquifer 80 m thick, horizontal and vertical permeability were equal to 5.2 x 10^{-12} \text{ m}^2, and longitudinal dispersivity length was set to 0.001 m. This latter value was intentionally set low so that dispersion effects would not mask any small differences that might occur between the results. The values for physical properties of fluid and aquifer are typical values for Hawai‘i ground-water systems and are discussed in greater detail in later sections (see regional modeling section and table 6.1). Both models used the same relationship to determine temperature-dependent viscosity. For most simulations, the time-step length initially started at 0.52 days and increased by a factor of 1.01 until a maximum time-step length of 6.3 days was attained. Some simulations used a constant time-step length of 0.52 days. A summary of all of the simulation options used is shown in table 5.1.

Run V1 is the simplest case, with constant fluid density and fluid viscosity based on a fluid temperature of 24°C. A constant upward flow gradient is maintained using specified pressure conditions at the top and bottom of the mesh, and the inflowing fluid required to maintain the specified pressure has a constant temperature of 95°C. The initial temperature is set to 24°C everywhere in the mesh except the base and top which are set to constant values of 95°C. The SUTRA numerical code had to be modified slightly
Table 5.1. Details of comparisons of energy-transport simulations between SUTRA and CFEST

<table>
<thead>
<tr>
<th>Run</th>
<th>Fluid Density</th>
<th>Fluid Viscosity</th>
<th>Thermal Conductivity and Fluid Specific Heat Capacity</th>
<th>Temperature Boundary at Top</th>
<th>Temperature Boundary at Base</th>
<th>Fluid Boundary at Top</th>
<th>Fluid Boundary at Base</th>
<th>Time Step Length</th>
</tr>
</thead>
<tbody>
<tr>
<td>V1</td>
<td>constant</td>
<td>constant</td>
<td>constant temperature @ 95°C</td>
<td>constant temperature @ 95°C</td>
<td>constant pressure with fluid @ 95°C</td>
<td>constant pressure with fluid @ 95°C</td>
<td>0.52-6.3 days</td>
<td></td>
</tr>
<tr>
<td>V1a</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>not specified</td>
<td>**</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V1b</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>energy flux</td>
<td>**</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V2</td>
<td>**</td>
<td>variable</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V3</td>
<td>variable</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V4</td>
<td>variable</td>
<td>variable</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V4a</td>
<td>variable</td>
<td>variable</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V5</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>energy flux</td>
<td>not specified</td>
<td>**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>V6</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>not specified</td>
<td>**</td>
<td>**</td>
<td>flux of 0.001 kg/s @ 95°C</td>
<td></td>
</tr>
<tr>
<td>V7</td>
<td>**</td>
<td>variable</td>
<td>**</td>
<td>not specified</td>
<td>**</td>
<td>**</td>
<td>flux of 0.001 kg/s @ 95°C</td>
<td></td>
</tr>
<tr>
<td>V7a</td>
<td>**</td>
<td>variable</td>
<td>**</td>
<td>not specified</td>
<td>**</td>
<td>**</td>
<td>flux of 0.001 kg/s @ 95°C</td>
<td>0.52 days</td>
</tr>
<tr>
<td>V8</td>
<td>variable</td>
<td>variable</td>
<td>**</td>
<td>not specified</td>
<td>**</td>
<td>**</td>
<td>flux of 0.001 kg/s @ 95°C</td>
<td></td>
</tr>
<tr>
<td>V8a</td>
<td>variable</td>
<td>variable</td>
<td>**</td>
<td>not specified</td>
<td>**</td>
<td>**</td>
<td>flux of 0.001 kg/s @ 95°C</td>
<td>0.52 days</td>
</tr>
</tbody>
</table>

Starting background temperature is 24°C for all simulations.
** - same as Run V1
to remove the temperature-dependent fluid viscosity function and replace it with a routine to maintain constant viscosity at all temperatures. The simulation progressed for 5,000 days and the temperature was monitored at two distances above the base, 8.4 and 37.9 m (fig. 5.8a). The CFEST simulation predicts a slightly quicker rise in temperature at the monitoring points than the SUTRA simulation does. Both simulations reach the steady-state temperature of 95°C as is expected for this scenario. These results suggest that CFEST tends to create a little more numerical dispersion than SUTRA for an identical simulation even at time-step lengths and element widths that are relatively small. Two similar simulations (Runs V1a and V1b) were completed to observe the effects of changing the temperature boundary condition at the top of the mesh on the energy transport. In the first case, the top temperature boundary was not specified at all, and in the second case it was specified as a source of energy flux without fluid. For both SUTRA and CFEST, unspecified boundary nodes are regarded as perfectly insulated boundaries where neither energy conduction nor dispersion can take place. Energy is only transported across the boundary advectively. Neither case changed the results at all and the temperature observations plot identically on the results of Run V1.

Run V2 was identical to Run V1 with the exception of the way fluid viscosity was modeled. In the SUTRA simulation, the built-in temperature-dependent viscosity function was used. In the CFEST simulation, a tabular look-up option estimated viscosity by linear interpolation between temperature-viscosity values entered in the input data set at 10°C intervals. The temperature-viscosity values entered into the input data set were calculated manually using the same temperature-dependent viscosity equation used in the SUTRA code. The results show that when the variable viscosity option is used, both
Figure 5.8 Comparisons between SUTRA and CFEST models, runs V1-V4.
the CFEST and SUTRA solutions reached steady-state about 50 percent faster than in Run V1 because the lower viscosity values increased the hydraulic conductivity of the aquifer (fig. 5.8b). The CFEST result predicted a slightly quicker rise in temperature 8.4 meters from the base of the mesh but inexplicably, a slower rise at the observation point 37.9 m from the base. The simulated temperature at this point was as much as 10°C cooler halfway through the simulation. Both models eventually predicted the same correct steady-state temperature of 95°C at the end of the simulation.

Run V3 was identical to Run V1 with the exception that temperature-dependent fluid density was included in the simulation. Both SUTRA and CFEST used a linear temperature-density relationship. When this option was employed, energy propagates upward through the aquifer almost ten times faster than when it was not (fig. 5.8c). Lower fluid density led to a lower hydraulic conductivity, but more importantly a density gradient was set up which allowed fluid to rise buoyantly carrying energy with it. The onset of free convection in this scenario should coincide with a Rayleigh number greater than $4\pi^2$ (Marsily, 1982), with the Rayleigh number defined as:

$$R_a = g \frac{\tau \rho (\rho c) k}{\mu \lambda} l \Delta T$$

(5.36)

where:

- $\tau$ = fluid coefficient of thermal volumetric expansion [°C⁻¹]
- $\rho$ = fluid density [M/L³]
- $c$ = fluid heat capacity [E/M·°C]
- $k$ = permeability [L²]
- $\mu$ = fluid viscosity [M/L·T]
- $\lambda$ = equivalent thermal conductivity of porous media [E/T·L·°C]
- $l$ = thickness of the layer [L]
- $\Delta T$ = temperature difference across layer [°C]
For this simulation the Rayleigh number is:

\[
(9.81)(5.206 \times 10^{-4})(996.9)(996.9)(4182)(5.2 \times 10^{-12})(80)(71) \approx 326
\]

This value is far above the convection threshold of \(4\pi^2 \approx 40\) so upward convective flow occurs more efficiently, causing the energy to be transported quickly and thus the numerical solution is more vulnerable to instabilities. Again, the CFEST simulation predicted a faster rise in temperature than the SUTRA simulation. In addition, the CFEST results showed some numerical oscillations before the steady-state temperature was reached. Because convection was occurring, the CFEST solution became somewhat unstable whereas the SUTRA simulation remained accurate.

Run V4 combined the changes in Runs V2 and V3 to simulate flow depending on both temperature-dependent viscosity and density. The combined effect means that fluid flow and energy transport occurred even faster, and the simulations reached a steady-state temperature over ten times faster than the case with constant values. Once again, CFEST predicted a more rapid rise in temperature than SUTRA did (fig. 5.8d). The numerical oscillations in the CFEST solution became more pronounced at both monitoring locations. Both models predicted the correct steady-state temperature of 95°C. An additional simulation (Run V4a) using CFEST was completed to test an optional way to model temperature-dependent fluid density and fluid heat capacity. This option uses a Taylor-series expansion equation (5.21) for describing fluid density as a function of temperature and a tabular look-up procedure for describing fluid heat capacity as a function of temperature. For this option to work, the base fluid-density value was set to
996.9 kg/m³, as the user's manual required. The results indicated that CFEST seems to work incorrectly when this option is chosen. The solution oscillates wildly and then settles down to a steady-state temperature that is obviously incorrect. Efforts to track down the cause of this error in the numerical code were not very rewarding. Therefore, no more solutions using this option were attempted.

Run V5 was similar to Run V1 but the boundary conditions were changed slightly. The top boundary condition was changed from a constant temperature condition to a dry energy flux condition which should have had no effect on the result as shown by the results of Run V1b. The bottom temperature boundary condition was unspecified, which meant that the base of the mesh would have an initial temperature of 24°C and would be heated up only by the flow of 95°C fluid entering at the base needed to maintain the specified pressure boundary. The results show that for this scenario, the CFEST and SUTRA results are very close to each other (Fig. 5.9a).

Run V6 was intended to simulate the effects of injection of hot fluid at the base of the aquifer which would lead to higher vertical transport velocities. The set up for this simulation was the same as for Run V1 with the exception of the fluid-flow boundary condition at the base which was changed from a specified pressure condition to a specified flux condition. A flux of 0.001 kg/s of 95°C fluid was added to the base of the mesh, a fluid-flux rate equal to 0.09 m³/d (≈ 0.02 gpm). This scenario produced energy-transport rates that are about 3-4 times faster than the V1 scenario (fig. 5.9b). The CFEST solution matched the SUTRA solution in the early times as it approached the steady-state result and then it deviated from the smooth SUTRA curve. The cause for the inaccuracies in the CFEST result are most likely due to time-step lengths that were too
Figure 5.9 Comparisons between SUTRA and CFEST models, runs V5-V8.
large although they were sufficiently small to allow SUTRA to produce an accurate result.

The next simulation, Run V7, was identical to Run V6 except that the temperature-dependent viscosity option was once again invoked. The results are much the same as the results for Run V6; initially the match is quite good but at larger time-step lengths the CFEST solution started to become inaccurate and underestimate the temperature at the observation point (fig. 5.9c). A CFEST solution with constant smaller time-step lengths (Run V7a) of 0.52 days was completed to see if the results would improve. For this simulation, the results did get better but they still deviated from the SUTRA results when higher temperatures were reached. Presumably, a match could be obtained by using even more finely-discretized time-step lengths but the amount of computer simulation time needed to produce a good match was considered excessive. Run V7a needed over 4,800 time steps and several days of computer time to complete. For future simulations eventually expected to cover up to 40 years of time, a time-step length of less than a half of one day is quite cumbersome.

The final set of simulations, Run V8, was the same as Run V6 with the exception that both temperature-dependent viscosity and density were modeled. This scenario is the closest to the actual conditions to be simulated in the ERZ study. The results show that CFEST is not robust enough to efficiently model the energy transport from an injection source. At identical time-step lengths to the SUTRA simulation, CFEST produces a wildly varying, inaccurate result (fig. 5.9d). When the time-step length is reduced to 0.52 days, the CFEST result is closer to the SUTRA result but still not acceptable. The steady-state temperature reached in the CFEST simulation is greater than 95°C, which is obviously an impossibility because the incoming fluid is only 95°C and no other energy sources are
included. Simulations using smaller time-step lengths were not completed because of the excessive length of time needed to get results and the impracticality of using a time-step length of less than a half a day for simulations covering tens of years.

Overall, SUTRA has proven to be a much more robust, accurate, and faster numerical code than CFEST for simulating energy transport. For identical fairly simple problems, both produce about the same result but SUTRA does so at least ten times faster. For more realistic scenarios, as flow and transport velocities increase, CFEST solutions deviate more from the correct value at identical time-step lengths to SUTRA solutions. Much smaller time-step lengths are needed for CFEST to approach the correct results and hence the total simulation time increases proportionally. Based on the results of the model verification simulations, CFEST was only used for solute transport simulations. However, density and viscosity values were both concentration and temperature dependent.
6. KILAUEA EAST RIFT ZONE NUMERICAL MODELS

In this chapter, several numerical models are used to investigate local and regional patterns of ground-water flow and energy and/or solute transport. The results of numerical simulations are compared to observed conditions to show that the models are valid. The models are then used to predict how changes in the aquifer fluid temperature and solute concentration due to geothermal development related activities might be observed in the existing well network and at coastal springs.

6.1. Modeling of Discharge From Shallow Fractures

6.1.1. Energy-Transport Modeling Approach

The three hottest and most complete temperature profiles (GTW-3, MW-2, and Malama Ki Well) were used in matching the energy-transport models of hot water discharge from a postulated fracture into the shallow ground-water flow system. No other shallow wells are deep enough to provide informative temperature profiles that could be utilized. This was based on a conceptual model of the local-scale flow system depicted in figure 6.1. Several simplifying assumptions were made in the conceptual model. These assumptions are that the well is down gradient from the discharging fracture and in the center of the hot water plume, that the fracture has been discharging hot water at least 40 years (since the 1955 magma injection and subsequent eruption), that both the
Figure 6.1 Conceptual model of discharging fracture.
discharging water and the ambient aquifer water are fresh, and that only a single-phase fluid would be considered.

6.1.2. Mesh Design and Boundary Conditions

A two-dimensional SUTRA mesh of 4,541-elements (fig. 6.2) was created for all of the plume simulations. Each element was 4.2105 m long, 5.0509 m deep, and 1 m wide (unit width) and the mesh represented a section of the aquifer 80 m deep and 1,200 m long. The top boundary was maintained as a constant fluid-flux boundary with recharge of 84 cm/yr at 24°C. All three of the wells lie in a zone reported by Takasaki (1993) to receive between 25 and 127 cm/yr of recharge. The bottom of the mesh was maintained as a no-flow boundary for fluid and a constant flux boundary for energy with 0.12 J/s added at each node. Kauahikaua (1993) presented calculated heat flow values for seven wells in the KERZ based on deep temperature logs. The range was between 370 and 820 mW/m² and the average was about 500 mW/m² (~ 0.5 J/s·m² x element area of 4.2105 m²). The right side of the mesh was maintained as a specified pressure boundary and the left was a specified fluid flux boundary to maintain a constant flow gradient from left to right consistent with the ambient flow gradient measured in the aquifer. A total of 0.152 kg/s of freshwater inflow was added to the left side of the mesh to be consistent with estimated ground-water flux in the ERZ. This volume is equivalent to ~215 Mgal/day through the 500 m deep, 10-km wide rift zone, and is based on the recharge estimates shown in figure 4.1 of Takasaki (1993). The discharging fracture was simulated as a column of fluid source nodes located 100 m from the left boundary of the mesh. The injected water in all cases
Figure 6.2  Finite-element mesh used in discharging fracture simulations.
was kept at a temperature of 100°C. The temperature of the water at the top of the aquifer cannot be greater than 100°C because at that pressure (atmospheric) the rest of the energy would be lost to steam above the water table (D. Thomas, pers. commun., 1994). Because SUTRA is not designed to handle steam flow or energy loss due to steam loss at the water table, higher temperatures could not be simulated accurately. The initial temperature of the aquifer was set at 24°C and a constant chloride concentration consistent with freshwater values was assumed for all simulations.

6.1.3. Input Parameters

Table 6.1 lists the values of fluid and solid matrix properties held constant in all of the fracture-discharge simulations. The values of fluid compressibility, fluid thermal conductivity, fluid specific heat, solute diffusivity, and the composition density ratio were obtained directly from the SUTRA manual. A manual titled Seawater and Seawater Distillation (Höming, 1978) provided the information on fluid viscosity, fluid thermal expansivity and the saltwater concentration-density ratio. The fluid density and internal energy values were obtained from published steam tables (Van Wylen and Sonntag, 1976). Fluid density ranges from 1,024 kg/m³ for saltwater at 20°C to 958 kg/m³ for freshwater at 100°C (fig. 6.3). Saltwater heated to 80°C has about the same density as freshwater at 20°C. Fluid viscosity ranges from 1.072 x 10⁻³ kg/(m·s) for saltwater at 20°C to 0.281 x 10⁻³ kg/(m·s) for freshwater at 100°C.

The solid matrix parameters were obtained from reported values used in other basalt aquifer modeling studies. Souza and Voss (1987) tested a range of matrix
Table 6.1. Fluid and solid matrix properties.

<table>
<thead>
<tr>
<th>Fluid Parameter</th>
<th>Value</th>
<th>Units</th>
<th>CFEST symbol</th>
<th>SUTRA symbol</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Compressibility</td>
<td>$4.47 \times 10^{-10}$</td>
<td>$m \cdot s^2/kg$</td>
<td>$C_p$</td>
<td>$\beta$</td>
<td>SUTRA manual</td>
</tr>
<tr>
<td>Thermal expansivity</td>
<td>-0.000519</td>
<td>$kg/m^3 \cdot ^\circ C$</td>
<td>$C_v$</td>
<td>$\frac{1}{\rho} \frac{\partial \rho}{\partial t}$</td>
<td>Hõmig</td>
</tr>
<tr>
<td>Composition density ratio</td>
<td>$1.3456 \times 10^{-6}$</td>
<td>$(mg/l)^{-1}$</td>
<td>$C_v$</td>
<td>$\frac{\partial \rho}{\partial C}$</td>
<td>Hõmig</td>
</tr>
<tr>
<td>Composition density ratio</td>
<td>700.</td>
<td>$kg/m^3$</td>
<td>$\rho$</td>
<td>$\rho$</td>
<td>SUTRA manual</td>
</tr>
<tr>
<td>Internal energy</td>
<td>83950.</td>
<td>$J/kg$</td>
<td>$U_e$</td>
<td>$U_e$</td>
<td>Van Wylen</td>
</tr>
<tr>
<td>Base density</td>
<td>996.9</td>
<td>$kg/m^3$</td>
<td>$\rho_e$</td>
<td>$\rho_e$</td>
<td>Van Wylen</td>
</tr>
<tr>
<td>Viscosity</td>
<td>0.001001</td>
<td>$kg/(m \cdot s)$</td>
<td>$\mu$</td>
<td>$\mu$</td>
<td>Hõmig</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>0.6</td>
<td>$J/(s \cdot m^\circ C)$</td>
<td>$E$</td>
<td>$E$</td>
<td>SUTRA manual</td>
</tr>
<tr>
<td>Specific heat capacity</td>
<td>4182.</td>
<td>$J/(kg \cdot ^\circ C)$</td>
<td>$C_p$</td>
<td>$C_p$</td>
<td>SUTRA manual</td>
</tr>
<tr>
<td>Diffusivity of solute</td>
<td>$1 \times 10^{-9}$</td>
<td>$m^2/s$</td>
<td>$\alpha$</td>
<td>$\alpha$</td>
<td>SUTRA manual</td>
</tr>
</tbody>
</table>

| Solid Matrix Parameter                   |                |                    |              |              |                  |
| Compressibility                          | $2.50 \times 10^{-9}$ | $m \cdot s^2/kg$  | $C_p$        | $\alpha$     | Souza and Voss (1987) |
| Thermal conductivity                     | 2.0           | $J/(s \cdot m^\circ C)$ | $E$          | $\lambda$    | Pruess et al. (1984) |
| Specific heat capacity                   | 900.          | $J/(kg \cdot ^\circ C)$ | $C_p$        | $C_p$        | Bodvarsson (1988)  |
| Density                                  | 2690.         | $kg/m^3$           | $\rho_v$     | $\rho_v$     | Johnson (1979)    |
| Porosity                                 | 0.05          |                    | $\theta$     | $\epsilon$   | Pruess et al. (1984) |
Figure 6.3 Relationship between fluid density and viscosity to concentration and temperature.
compressibility values over three orders of magnitude and determined that $2.5 \times 10^{-9}$ m·s$^2$/kg was the best value in their southern Oahu simulations. The values of solid matrix thermal conductivity and porosity were obtained from a modeling study of a geothermal field in a basalt aquifer in Iceland (Pruess et al., 1984). The solid matrix heat capacity was obtained from a similar Icelandic basalt aquifer modeling study reported by Bodvarsson (1988). Density of the solid matrix is based on the reported values of dry bulk density obtained from the upper 680 m of the HGP-A core hole (Johnson, 1979).

For the fracture simulations near GTW-3 and MW-2, the horizontal permeability of the aquifer was held constant at $4.7372 \times 10^{-11}$ m$^2$/s, a value equivalent to a hydraulic conductivity of 40 m/d for freshwater at 20°C. This value was judged to be a good representation of the rift zone hydraulic conductivity based on the tidal-response simulations and the regional simulations to be discussed later in this report. For the Malama Ki Well simulations, an additional value of horizontal conductivity for the aquifer outside of the rift zone was chosen to be equivalent to 400 m/d. Discussion about the assumptions and reasons for representing the flow conditions will be handled separately for each well.

6.1.4. Discussion and Results of Discharging-Fracture Simulations

GTW-3

GTW-3 is the hottest shallow well in the LERZ, and seems to be fairly close to a
fracture supplying a significant volume of hot water or steam from the deeper geothermal reservoir. It lies about 1 km east of the current geothermal production area and is most likely down gradient from the field (Thomas, 1994). The geologic map of the area (Moore and Trusdell, 1991) shows at least three mapped fractures that are visible at the land surface up gradient within 500 m of the well, and there are undoubtedly more which may not continue all the way up to the ground surface. The observed temperature profile shows that the hot water is getting right to the top of the aquifer and has not mixed very much with the cold meteoric water. The profile shows a drop in temperature of 50°C in the top 30 m of the aquifer, a steep gradient of 1.7°C/m.

The matching procedure involved getting the maximum temperature of the plume to match as well as the thickness of the lower part of the plume. An iterative procedure was used by first estimating a volume of discharging water and then the distance between the postulated fracture and the well. The best shaped profile was obtained when the majority (about 63 percent) of the hot water was added to the top two nodes of the mesh and the 18 lower nodes received the rest. From the geologic map, it was apparent that the distance between the well and the source was not likely to be more than 500 m and probably much less than that. Although there are many combinations of water volume and well distance that could theoretically produce the same profile, it was fairly easy to find a range of values that seemed reasonable.

Figure 6.4 shows the observed and simulated steady-state temperature profiles for the best fitting case. These results are for the case where the well is about 170 m down gradient from a fracture producing $1.26 \times 10^{-1}$ kg/s of 100°C water for 40 years, which is equivalent to about 11.37 m$^3$/d.
Figure 6.4 Observed and simulated temperature profiles, GTW-3.

- Simulation with well 170 m down gradient from a fracture producing 11.4 cu. m/d of 100 degree C fluid.

- Observed temperature profile GTW-3.
The enthalpy of pure 100°C water is 419.04 kJ/kg (Van Wylen and Sonntag, 1976). The simulated fracture would therefore be adding energy to the aquifer at the rate of 52.8 kJ/s. This energy-flux rate can be used to estimate the volume of water or steam of a different temperature that would add an equivalent amount of energy to the aquifer. For example, an equivalent amount of energy would be added by a fracture discharging pure steam at 100°C at a rate of \(2.11 \times 10^{-2}\) kg/s, but a much larger volume of 3,050 m\(^3\)/day. A fracture discharging 200°C steam would do so at a rate of \(1.9 \times 10^{-2}\) kg/s (or 208 m\(^3\)/day). These volumes are calculated for a fracture that is a unit length of 1 m. A fracture 500 m long would produce 5,685 m\(^3\)/day (~ 1.5 Mgd) of 100°C water. The total energy flux for a fracture of this length would be:

\[
\frac{52.8 \text{ kJ}}{s} \times \frac{500 \text{ m}}{1 \text{ m}} \times \frac{1 \text{ kW} \cdot \text{s}}{1 \text{ kJ}} \times \frac{MW}{10^3 \text{kW}} = 26.4 \text{ MW}
\]

(6.1)

The temperature profile at this location does not change from about four years into the simulation until the end of the simulation at 40 years. So it is essentially at a "steady-state" condition for about 30 years. The temperature profiles collected 20 years ago in 1974-1976 are almost the same as the most recent profile suggesting that the assumption about the minimum age of the source fracture is reasonable. It is also reasonable to suggest that the fracture-discharge rate has not changed much for 40 years.
MW-2

MW-2, possibly the most significant in terms of monitoring the current geothermal production area due to its position immediately adjacent to the area of interest, is also fairly close to a fracture supplying hot water or steam from the deeper geothermal reservoir. This well lies less than 1 km from the current geothermal production wells KS-9 and KS-10. The geologic map of the area shows several mapped fractures that are up gradient and in the vicinity of the well. More importantly, this well appears to be within 150 m of the surface intersection of the KS-8 Well fracture mapped by Teplow Geologic (written commun., 1994) based on well logs from KS-7, 8, 9, and 10 Wells and CSAMT profiling. This fracture is reported to strike roughly parallel to the ERZ (N63E) and have a near-vertical dip (82°). Currently, 25 MW of power are generated from Wells KS-9 and KS-10 which are both thought to tap the KS-8 Well fracture. Much like the GTW-3 profile, the MW-2 observed temperature profile shows that the hot water is getting right to the top of the aquifer and has not mixed very much with the cold meteoric water. The profile shows a drop in temperature of 30°C in the top 20 m of the aquifer, also a fairly steep gradient of 1.5°C/m.

The same fluid source distribution as described for GTW-3 was used for the MW-2 simulations. Figure 6.5 shows the observed and simulated steady-state temperature profiles for the best fitting case. The profile in this simulation also reached a "steady-state" condition after about 3 or 4 years. This scenario is for the case where the well is about 105 m down gradient from a fracture producing $2.545 \times 10^{-2}$ kg/s ($2.3 \text{ m}^3/\text{d}$) of 100°C water. The simulated fracture would therefore be adding energy to the aquifer at the rate
simulation with well 105 m down gradient from a fracture producing 2.3 cu. m/d of 100 degree C fluid.

Figure 6.5 Observed and simulated temperature profiles, MW-2.
of 10.7 kJ/s. An equivalent amount of energy would be provided by $3.99 \times 10^{-3}$ kg/s (576 m$^3$/d) of 100°C steam or $3.82 \times 10^{-3}$ kg/s (42 m$^3$/d) of 200°C steam. Proprietary geophysical mapping interpretations by Teplow Geologic (written commun., 1994) show that the fracture may be at least 500 m long and probably longer. A fracture 500 m long would produce $1,150 \text{ m}^3/\text{day} (-0.3 \text{ Mgd})$ of 100°C water, and the total energy output would be about 5.3 MW.

**Malama Ki Well**

This well is located about 600 m south of the surface expression of the ERZ so the horizontal hydraulic conductivity of the aquifer between the fluid source and the well is probably higher than for the wells located inside the rift zone. The geologic map of the area shows one mapped fracture that is up gradient and in the vicinity of the well. But it is felt that the source of hot water observed in the Malama Ki Well is associated with Puulena Crater, a large phreato-magmatic eruptive feature on the southern edge of the LERZ. This crater is over 100 m deep and reportedly still contains some active steam vents near the bottom (F. Trusdell, USGS, oral commun.). There is no historic record of the eruption associated with this feature but it must have been quite impressive to have produced such a large crater. Currently, USGS personnel are attempting to map the extensive tuff deposits from this vent. These data will provide more insight into the mechanisms involved in this event which probably involved the injection of magma into a reservoir of ground water. A value for hydraulic conductivity of 400 m/d was chosen as an intermediate value between the rift zone conductivity (40 m/d) and the flank conductivity
(4,000 m/d) because the area of interest lies in the zone of transition between the rift and the flank where dikes are most likely present but not abundant. The profile shows the same pattern of hot water floating on cooler water but the hottest part of the profile is deeper because of increased mixing as the plume migrates down gradient. The well is not deep enough to provide a good profile of the hot water plume below an altitude of -12 m. The same fluid source distribution as described discussed previously was used for the Malama Ki Well simulations.

Figure 6.6 shows the observed and simulated steady-state temperature profiles for the best fitting case. This result is for the case where the well is about 730 m down gradient from a fracture producing 0.061 kg/s (5.5 m³/d) of 100°C water. The simulated fracture would be adding energy to the aquifer at the rate of 25.6 kJ/s. An equivalent amount of energy would be provided by 9.55 x 10⁻³ kg/s (1381 m³/d) of 100°C steam or 9.15 x 10⁻³ kg/s (101 m³/d) of 200°C steam. A fracture 500 m long would produce 2,750 m³/day (~ 0.73 Mgd) of 100°C water, although the nature of the Puulenena eruption suggests that the fluid source could be supplied through a more pipe-like conduit. The total energy output for a 500-m long fracture would be about 12.8 MW. The temperature profile reached a "steady-state" condition after about 12 years of simulated time, due to its longer distance from the discharging fracture.

Transient Temperature Monitoring

The three discharging-fracture simulations are important tools for predicting how up-gradient temperature changes will affect the temperature measured in the monitoring
simulation with well 730 m down gradient from a fracture producing 5.5 cu. m/d of 100 degree C fluid

observed temperature profile
Malama Ki Well

Figure 6.6 Observed and simulated temperature profiles, Malama Ki Well.
wells. The temperature distribution after 40 years of simulated time was used as the starting point for a series of simulations in which the fracture-discharge rate was varied to simulate either a change in fracture characteristics or the addition of another source of hot water. This is analogous to the situation where a leaky geothermal well or injection well is present up gradient from the monitoring well. The fracture-discharge rate was scaled by the following amounts: 2, 1.5, 1.25, 0.75, 0.5, and 0, and each change was instantaneous. Each simulation was allowed to continue for 40 more years and the change in temperature over time at the top of the aquifer and the temperature profile at the end of the simulation was recorded. Table 6.2 is a compilation of all of the transient scenarios and resulting simulated temperature changes.

Figure 6.7 shows the continuous temperature values at the top of the water table for each simulation plotted until a new "steady-state" condition was reached. The results are not symmetric around the original "steady-state" temperature because the temperature of the fracture water is held at 100°C. This serves as a limiting factor on the rise of the water temperature down gradient at the well location. For the case when the flow from the fracture is completely stopped, it takes about 4.5 years for the simulated profile in GTW-3 to return to background temperatures, 2.5 years for MW-2 and about 14 years for Malama Ki Well.

Figure 6.8 shows the "steady-state" temperature profiles for each simulation. The maximum temperature in GTW-3 increased by only about 8°C when the simulated fracture-discharge rate was doubled. Whereas, the maximum temperature in MW-2 and Malama Ki Well increased by 18 and 15°C, respectively, for the doubled rate. Again, this is because of the limit on the maximum temperature allowed by the condition at the
Table 6.2. Transient fracture-discharge simulation scenarios.

<table>
<thead>
<tr>
<th>Well</th>
<th>Distance from fracture to well (meters)</th>
<th>Fracture-discharge rate (cu. meters/day)</th>
<th>Equivalent energy (kJ/s)</th>
<th>Time for temperature to change 3°C (days)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>GTW-3</td>
<td>170</td>
<td>11.4</td>
<td>52.2</td>
<td>-</td>
<td>base case</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>0</td>
<td>0</td>
<td>105</td>
<td>no fracture flow</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>5.7</td>
<td>26.1</td>
<td>182</td>
<td>50% reduction</td>
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<tr>
<td>*</td>
<td>*</td>
<td>8.6</td>
<td>39.2</td>
<td>517</td>
<td>25% reduction</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>14.3</td>
<td>65.3</td>
<td>&gt; 2000</td>
<td>25% increase, did not reach a 3°C increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>17.1</td>
<td>78.3</td>
<td>215</td>
<td>50% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>22.8</td>
<td>104.4</td>
<td>74</td>
<td>100% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>34.2</td>
<td>156.6</td>
<td>25</td>
<td>300% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>57.0</td>
<td>261.0</td>
<td>14</td>
<td>500% increase</td>
</tr>
<tr>
<td>MW-2</td>
<td>105</td>
<td>2.3</td>
<td>10.7</td>
<td>-</td>
<td>base case</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>0</td>
<td>0</td>
<td>131</td>
<td>no fracture flow</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>1.2</td>
<td>5.4</td>
<td>160</td>
<td>50% reduction</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>1.7</td>
<td>8.0</td>
<td>193</td>
<td>25% reduction</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>2.9</td>
<td>13.4</td>
<td>&gt; 2000</td>
<td>25% increase, did not reach a 3°C increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>3.8</td>
<td>16.1</td>
<td>191</td>
<td>50% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>4.6</td>
<td>21.4</td>
<td>109</td>
<td>100% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>13.7</td>
<td>63.7</td>
<td>38</td>
<td>600% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>25.1</td>
<td>116.8</td>
<td>23</td>
<td>1,100% increase</td>
</tr>
<tr>
<td>*</td>
<td>*</td>
<td>47.9</td>
<td>222.8</td>
<td>12</td>
<td>2,100% increase</td>
</tr>
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Table 6.2. Transient fracture-discharge simulation scenarios - cont.

<table>
<thead>
<tr>
<th>Well</th>
<th>Distance from fracture to well (meters)</th>
<th>Fracture-discharge rate (cu. meters/day)</th>
<th>Equivalent energy (kJ/s)</th>
<th>Time for temperature to change 3°C (days)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Malama Ki</td>
<td>730</td>
<td>5.5</td>
<td>25.6</td>
<td>-</td>
<td>base case</td>
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<td></td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>527</td>
<td>no fracture flow</td>
</tr>
<tr>
<td></td>
<td>2.8</td>
<td>12.8</td>
<td>712</td>
<td></td>
<td>50% reduction</td>
</tr>
<tr>
<td></td>
<td>4.1</td>
<td>19.2</td>
<td>970</td>
<td></td>
<td>25% reduction</td>
</tr>
<tr>
<td></td>
<td>6.9</td>
<td>32.0</td>
<td>880</td>
<td></td>
<td>25% increase</td>
</tr>
<tr>
<td></td>
<td>8.3</td>
<td>38.4</td>
<td>560</td>
<td></td>
<td>50% increase</td>
</tr>
<tr>
<td></td>
<td>11.0</td>
<td>51.2</td>
<td>393</td>
<td></td>
<td>100% increase</td>
</tr>
<tr>
<td></td>
<td>16.9</td>
<td>78.7</td>
<td>241</td>
<td></td>
<td>310% increase</td>
</tr>
<tr>
<td></td>
<td>28.3</td>
<td>131.7</td>
<td>143</td>
<td></td>
<td>510% increase</td>
</tr>
<tr>
<td></td>
<td>51.1</td>
<td>237.8</td>
<td>76</td>
<td></td>
<td>930% increase</td>
</tr>
</tbody>
</table>
Figure 6.7 Simulated transient temperature changes in GTW-3, MW-2 and Malama Ki Well.
Figure 6.8 Simulated temperature profiles after 40 years in GTW-3, MW-2 and Malama Ki Well.
water-table surface. GTW-3 is already near the upper temperature limit so it cannot get much hotter no matter how much more hot water is added. When the simulated flow volume is cut in half, all three wells show a temperature decline of 13-15°C.

From continuous monitoring of the wells, it was found that the ambient temperature in the wells can vary about 3°C just due to daily changes in recharge infiltrating into the aquifer from above. Therefore, any temperature change would have to be greater than about ±3°C to be noticeable above the ordinary ambient temperature fluctuations. Figure 6.9 also shows the time-temperature data but the axes are scaled to just show a 3°C rise above the initial temperature. At GTW-3, it took about 70 days for the simulated well temperature to rise 3°C after the fracture-discharge rate was doubled. This corresponds to an addition of 11.4 m³/d (105.6 kJ/s) to the shallow aquifer. Thomas (oral commun., 1995) suggests that in the case of a catastrophic failure of the well casing, a geothermal well would release a few gallons per minute of hot water into the shallow aquifer if the pressure inside the casing was higher than the hydrostatic pressure in the aquifer. A two gallons per minute flux rate is equivalent to 11.4 m³/d of 100°C water. The failure should be detected almost immediately by monitoring equipment in the geothermal well casing. Therefore, it is expected that release of water to the aquifer would not last longer than a day or two. These fracture-flow simulations can be considered a "worst" case scenario when fluid leaks constantly for weeks. Simulations were also run for the case when the additional discharge to the aquifer totaled 22.8 and 45.6 m³/d. These new rates caused a rise of 3°C after about 25 and 15 days, respectively.

When the simulated discharge rate up gradient of MW-2 is increased by 11.4 m³/d of 100°C water it took about 40 days for the temperature to rise 3°C. Additions of 22.8 and
Figure 6.9 Simulated transient temperature changes in GTW–3, MW–2 and Malama Ki Well.
45.6 m$^3$/d caused changes in about 22 and 12 days. Although Malama Ki Well is not directly down gradient from the current geothermal development area, it is still interesting to investigate the effects of fracture-volume changes. For an addition of 11.4 m$^3$/d, it took about 240 days for the temperature to rise 3°C. Additions of 22.8 and 45.6 m$^3$/d caused a 3°C rise at the well location in about 240 and 140 days.

**Transient Concentration Simulations**

Thomas (1994) suggested that a good indicator for the presence of geothermal water in MW-2 would be the sulfate ion. Sulfate ion would be produced by the oxidation of H$_2$S entering the shallow ground water from either a deep fracture or a geothermal fluid reinjection well casing leak. He estimates that the concentration of sulfate would have to increase by about 80 mg/kg above the background concentration of about 100 mg/kg at MW-2 to show an externally induced change in sulfate composition.

Several SUTRA solute-transport simulations were run using the same mesh and input parameters that were used for the energy-transport model (table 6.3). The fluid density and viscosity were based on a temperature of 70°C, the approximate average temperature of the aquifer between the fracture and the well location in the previous simulation. The background sulfate concentration was set at 4 mg/kg based on measured sulfate concentrations at the Paradise Park Well. For the first simulation, the fracture was discharging at a flow rate of 2.3 m$^3$/d, the rate determined from the energy-transport simulations. The sulfate concentration of the water discharging from the fracture was adjusted until the sulfate concentration at the MW-2 location was 100 mg/kg. The best
Table 6.3. Scenarios of transient concentration transport from fracture discharge.

<table>
<thead>
<tr>
<th>Well</th>
<th>Distance from fracture to well (meters)</th>
<th>Fracture-discharge rate (cu. meters/day)</th>
<th>Equivalent sulfate flux (mg/s)</th>
<th>Time for concentration to increase 80 mg/kg (days)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>MW-2</td>
<td>105</td>
<td>2.3</td>
<td>7.2</td>
<td>-</td>
<td>base case</td>
</tr>
<tr>
<td>-</td>
<td>-</td>
<td>4.1</td>
<td>12.8</td>
<td>60</td>
<td>minimum value needed to reach 80 mg/kg threshold</td>
</tr>
<tr>
<td>-</td>
<td>-</td>
<td>13.7</td>
<td>42.9</td>
<td>13</td>
<td>600% increase</td>
</tr>
<tr>
<td>-</td>
<td>-</td>
<td>25.1</td>
<td>78.6</td>
<td>9</td>
<td>1,100% increase</td>
</tr>
<tr>
<td>-</td>
<td>-</td>
<td>47.9</td>
<td>149.9</td>
<td>5</td>
<td>2,100% increase</td>
</tr>
</tbody>
</table>
result was obtained for a fracture-discharge sulfate concentration of 275 mg/kg (table 6.3). It took about 12 days of simulated time for the concentration at the MW-2 location to begin to increase and about two months to reach a "steady-state" condition. This "steady-state" concentration distribution was used as a starting point for the subsequent simulations. The first case was where the fracture-discharge rate was increased by 1.8 m$^3$/d to 4.1 m$^3$/d and the concentration at MW-2 was increased by 80 mg/kg, the proposed threshold for determining an externally induced change in sulfate composition. At this new fracture-discharge rate, it took about nine days for the observed sulfate concentration at MW-2 to start to increase, and about 60 days to reach a "steady-state" concentration which was 80 mg/kg above the starting concentration. Additional discharge volumes of 11.4, 22.8, and 45.6 m$^3$/d were added and the changes were again recorded (fig. 6.10). These simulations showed sulfate-concentration increases of at least 80 mg/kg at MW-2 in 13, 9, and 5 days from the change in the simulated discharge rate.

6.1.5. Model Sensitivity

One way to determine what physical and numerical processes control the results of a simulation is to perform a sensitivity analysis of the parameters used in setting up the model. This was performed for seven parameters used in the previously discussed transient energy and solute-transport simulations: rock density (energy only), fluid viscosity (solute only), porosity, anisotropy ratio, permeability, longitudinal dispersivity length, and transverse dispersivity length. For the energy-transport sensitivity analysis, each simulation started with a constant temperature distribution of 24°C and then 2.3 m$^3$/d
Figure 6.10 Simulated transient sulfate concentration changes in MW-2.
of 100°C water was added to the mesh. For the solute-transport sensitivity analysis, each simulation started with a constant sulfate distribution of 4 mg/kg and 2.3 m$^3$/d of water containing 275 mg/kg of sulfate was added to the mesh. A base-case simulation was performed first with all of the parameters maintained at values felt to be the most appropriate. Subsequent simulations were performed after changing a single parameter to a higher or lower value, usually scaled by an order of magnitude. Each simulation continued for 1.5 years and the transient rise in temperature or sulfate concentration and a temperature profile at the end of the simulated time 100 m from the source were observed (figs. 6.11a-k and 6.12).

Rock density is important in transient energy-transport simulations because it occurs in the time-dependent term in the energy-transport equation where it affects the storage of energy in the solid matrix. For simulations continued to steady-state conditions, the rock density will not affect the final result. The range of rock density values used in the sensitivity analysis was obtained from the reported values of dry bulk density measured in the HGP-A core which was 2,350 to 2,910 kg/m$^3$ (Johnson, 1979). In the simulation, when the rock density was lowered, the energy storage capacity of the aquifer was lowered and energy was transported faster through the aquifer as indicated by the rise in temperature at the down-gradient monitoring point sooner than in the base-case simulation. In the base-case simulation, it took about 160 days for the observed temperature to rise to 30°C and in the lower-rock-density simulation, it took about 140 days (fig. 6.11a). This represents a 14 percent faster travel time for a 14 percent lower rock density. Conversely, an eight percent higher value of rock density results in an eight percent slower propagation of the energy plume. All three simulations will eventually reach the same
Figure 6.11 Sensitivity analysis for discharging fracture simulations, transient temperature results.
Figure 6.11 - cont. Sensitivity analysis for discharging fracture simulations, transient concentration results.
Figure 6.12 Sensitivity analysis for discharging fracture simulations, temperature profiles.
steady-state temperature distribution as shown by the almost identical temperature profiles at the end of each simulation (fig. 6.12a).

Aquifer porosity occurs in the fluid flow equation and the energy and solute transport equations and affects both transient and steady-state results. In the flow equation it affects the velocity at which fluid travels through the aquifer and the storage of water in the pore spaces. In the energy-transport equation, it affects the effective energy storage capacity and the effective thermal conductivity of the aquifer as well as the advective and dispersive flow characteristics. In general, a lower porosity should lead to a lower energy storage capacity and a higher energy-transport velocity in the aquifer. One shortcoming of the numerical energy-transport models is that they do not differentiate between the true porosity which includes all pore space and the effective porosity which just includes interconnected pores that control water movement. The effective energy storage capacity and effective thermal conductivity of the aquifer should be based on the true porosity while the transport velocity is based on the effective porosity. But as the energy-transport sensitivity analysis shows, for the range of porosity values likely in a basalt aquifer, the differences are insignificant (figs. 6.11b and 6.12b).

In energy simulations, the time it takes for the temperature at the monitoring point to increase to 30°C using a porosity of 0.01 is only about 3 percent faster than for the base-case simulation using a porosity of 0.05. Raising the porosity to 0.1 slowed the simulated transport time by 3 percent. By the time the temperature increases to 50°C, there is only a 1 percent difference in transport time between the different simulations. Therefore, the energy simulations are not very sensitive to the value of porosity even though it occurs in most of the components of the numerical solution. Solute transport is
highly sensitive to changes in porosity. In simulations, when the porosity was lowered by 80 percent from 0.05 to 0.01, the time for the sulfate concentration at the observation point to begin to increase was about 80 percent shorter (fig. 6.12h). When the porosity was raised by 100 percent, the time increased by about 100 percent as well. All three simulations reached the same final steady-state concentration.

Anisotropy ratio is a measure of the difference between the horizontal and vertical permeability of the aquifer. In the fluid flow equation, permeability affects the rate at which fluid flows through the aquifer. In the base-case simulations, both values are equal giving a anisotropy ratio of 1:1. To get anisotropy ratios of 10:1 and 1:10, the vertical permeability was varied one order of magnitude lower and higher and the horizontal permeability was maintained at the original value of $4.7372 \times 10^{-11}$ m$^2$/s. In the energy-transport simulations, changing the anisotropy ratio to 10:1 causes a delay in the onset of the temperature rise of about 14 percent (fig. 6.11c). When the ratio is changed to 1:10, the onset of the temperature rise is about six percent sooner. The oscillation in the results at about 0.6 years for the 10:1 case is probably due to increased instability in the numerical solution introduced by changing the vertical permeability value by an order of magnitude. At later times, all three solutions approach the same temperature. The profile at the end of the simulation shows that the anisotropy ratio does not effect the spreading of the plume vertically in the aquifer (fig. 6.12c). This is probably because the simulations are set up to model mainly horizontal transport with only minor vertical flow or transport. In the solute transport simulations, changing the anisotropy ratio had only a minor effect on the final results. For the 10:1 case, the results are very close to the base case. For the 1:10 case, the simulated sulfate concentration is slightly lower because the
higher vertical permeability allows more vertical spreading and dilution of the solute plume.

The longitudinal dispersivity length is essentially a fitting parameter which occurs in the dispersive component of the energy and solute-transport equations. It affects the spreading of energy or solute in the direction parallel to the advective flow of the fluid transporting it. It is not a parameter such as porosity or rock density which is easily measured in the field or laboratory. So the basis for selecting an acceptable value of dispersivity length is much more subjective. Reported values of dispersivity length extend over several orders of magnitude depending on the field scale of the study performed. Typically, energy and solute transport dispersivity lengths differ for the same area of study. The selection of dispersivity length values for this study area (KERZ) is discussed in more detail in Chapter 6.3.1. For energy transport, the value was scaled one order of magnitude above and below the originally chosen value of 1 m. As shown by the results, lowering the value to 0.1 m has little effect on the transport time of the energy plume (figs. 6.11d and 6.12d). In the simulation, the observed rise in temperature takes slight longer to begin but it reaches a final temperature more quickly. This is because the lower dispersivity length causes less spreading and the energy-transport process is closer to slug flow. For the case where the dispersivity length is increased to 10 m, the initial rise in temperature occurs about 40 percent sooner but the overall rise in temperature is slower. The simulation was not carried out long enough for the temperature to approach the base-case temperature but it eventually should reach that value. These simulations suggest that the results of energy-transport simulations will be about the same for longitudinal dispersivity lengths less than 1 m. A longitudinal dispersivity length
approaching 10 m will produce a much slower transport time after the initial stages of the simulation. But for the case where energy transport is observed only 100 m from the source, a 10-m dispersivity length is not a realistic value to use. So in the range of realistic values, the transient energy-transport results are not exceptionally sensitive to the longitudinal dispersivity length.

Solute-transport simulations are more sensitive than energy-transport simulations to changes in longitudinal dispersivity length. In simulations, when the value of longitudinal dispersivity is lowered by one order of magnitude, the time is about 30 percent longer for the onset of solute-concentration increase (fig. 6.11j). But the whole slug of solute reaches the observation point sooner. The results demonstrate where numerical inaccuracies can occur when a too-small value of dispersivity is used. The numerical model has some "overshoot" before it settles down to the final steady-state value. For a large value of longitudinal dispersivity length, the rise in concentration is spread out a lot more and a lower steady-state value is reached. But for the case where solute transport is observed only 100 m from the source, a 50-m dispersivity length is not a realistic value to use.

Transverse dispersivity length is another fitting parameter which describes the spreading of energy or solute perpendicularly to the direction of fluid flow. For this analysis the value was scaled one order of magnitude above and below the originally chosen value of 0.1 m. As shown by the results, lowering or raising the value has little effect on the time of the initial rise in temperature or sulfate concentration at the monitoring point (fig. 6.11e). It does, however, affect the spread of the plume vertically in the aquifer (fig. 6.12e). Because, in these simulations, advective transport occurs mainly in the
horizontal direction, transverse dispersivity affects spreading mainly in the vertical direction. In the simulation where the transverse dispersivity length is lowered, the value at the monitoring point gets higher because the plume is not spread out as much as in the base case. The same amount of energy or solute is contained in a smaller area. This is evident in the temperature-profile plot. When the transverse dispersivity length is increased, the observed value is lower because the plume has been spread out over a larger area. At steady-state conditions, the temperature profile for each simulation will be different. The simulations show that the transverse dispersivity length, like the longitudinal dispersivity length, is less sensitive at the lower range of values tested (0.01 - 0.1 m for energy, 0.05 - 0.5 m for solute). Values in this range are more realistic representations of the dispersive processes because of the scale of the area of study.

For the discharging-fracture simulations, it may be wise to set the values of dispersivity length as low as possible without causing numerical instabilities in the solution. Bear (1972) states that the coefficient of longitudinal thermal dispersion (E_l) is negligible as compared with the coefficient of thermal diffusivity (\(\lambda / \rho c_p\)) at a Peclet number (Pe) less than 3,000 and in this range E_l = 0. The Peclet number for thermal transport is defined (Marsily, 1986) as:

\[ Pe = \frac{v l}{\lambda / \rho c_p} \]  

(6.3)

where \(l\) is the mean diameter of the grains in the solid matrix, \(v\) is the Darcy velocity and the other coefficients are defined previously. For the discharging-fracture simulations, \(v = 2.6 \times 10^{-5} \text{ m/s}, \lambda / \rho c_p = 7.7 \times 10^{-7} \text{ m}^2/\text{s}\), and \(l\) can be conservatively estimated at 1 m. This produces a \(Pe = 33.8\), well below the limit where thermal dispersion is considered
significant. Therefore, using a large (or any?) value of dispersivity length will introduce dispersion in the numerical solution while in reality, dispersivity is not a significant process in the transport of energy for these field conditions.

Fluid viscosity is present in the fluid flow equation in the conductivity term, which is a controlling factor of the rate of fluid flow through the aquifer. It is temperature and solute dependent; higher temperature and/or lower concentration fluid has a lower viscosity. For the low sulfate concentrations used in these simulations the dependence of viscosity on concentration changes is insignificant. Because the solute-transport sensitivity is performed at a constant temperature of 70°C, only one value of viscosity is used in the solution. The insensitivity associated with using a single viscosity value can be seen in this analysis. Values of viscosity for water at the ambient background temperature (0.0010 kg/m·s at 24°C) and at the maximum fracture-discharge temperature (0.0003 kg/m·s at 100°C) were used. The results show virtually no difference over the range of viscosity values tested (figs. 6.11f and 6.12b).

Fluid density is present in the fluid flow equation where it affects the rate of fluid flow through the aquifer. Fluid density is temperature and concentration dependent; higher densities occur with lower temperatures and/or higher solute concentrations. For low sulfate concentrations, like those used in these simulations, the dependence of density on concentration changes is insignificant. Because the solute-transport sensitivity is performed at a constant temperature of 70°C, only one value of density is used in the solution. The errors associated with using a single density value can be seen in the sensitivity analysis (fig. 6.12g). Values of density for saltwater at the ambient background temperature (1,024 kg/m³ at 24°C) and freshwater at the maximum fracture-discharge
temperature (958 kg/m$^3$ at 100°C) were used. The results show a linear dependence on fluid density. With a five percent higher fluid density than the base case (978 kg/m$^3$), solute transport is five percent slower than the base case.

6.2. Modeling of Shallow Well Continuous-Monitoring Records

6.2.1. Modeling Approach

'Water-level and temperature records from the Paradise Park Well, the reference well in the geochemistry monitoring program, MW-2, the closest down-gradient shallow well to the geothermal production area, and the Malama Ki Well, just south of the ERZ, were used to investigate the relationship between recharge, water level, and temperature, and to further investigate the hydraulic properties of the aquifer. The continuous water-level records from the Paradise Park Well contain several rapid water-level rise and decay features that correspond to extremely heavy rainfall events (fig. 4.9). SUTRA was used to simulate the transient response of water level in the aquifer to these recharge pulses. The fracture-discharge models discussed in the previous section were used to study the effects of fresh, cool recharge to the temperature distribution around MW-2 and the Malama Ki Well. And finally simulations of sulfate ion transport were run to investigate the effects of recharge on the dissolved-solids concentration of fluid in MW-2.
6.2.2. Paradise Park Well Simulations

Model Setup

A 3,625-element rectangular SUTRA mesh (fig. 6.13) was created for the transient Paradise Park Well simulations. The mesh represented a section of the aquifer 650 m deep and 11,500 m long, with each element being 100 m long and 5-65 m deep wide. The bottom of the mesh was maintained as a no-flow boundary for fluid. The right side of the mesh was maintained as a specified pressure boundary, and the left side was a specified fluid flux boundary to maintain a constant flow gradient from left to right consistent with the ambient flow gradient measured in the aquifer. The top boundary was maintained as a fluid-flux boundary with time-varying recharge. The recharge to the mesh was based on annual recharge estimates by Takasaki (1993) and recorded rainfall measurements at Old General Lyman Field in Hilo, about 20 km from the well site. From the map on p. 9 of Takasaki, recharge was chosen to be 127 cm/yr (50 in/yr). Daily rainfall records from Hilo (NOAA, 1992) were used to create an indexing function to convert the yearly rate into a daily value. Four months of daily rainfall totals were added together to get a rainfall total for September through December, 1992. Then each daily rainfall value was divided by the total to get a scale factor relating the daily value to the overall total. The recharge was then determined by multiplying the scale factor for each day by 127 cm/yr. So the final recharge used in the model was based on the estimated recharge at Paradise Park but the rainfall frequency pattern at Hilo. Daily rainfall records were also available from the other sites in the ERZ but they were not nearly as reliable as the NOAA data.
Figure 6.13 Finite-element mesh used in Paradise Park Well simulations.
Table 6.1 lists the values of fluid and solid matrix properties held constant in all of the transient-flow simulations. Fluid density and fluid viscosity values were for freshwater at 20°C. The starting horizontal permeability of the aquifer was based on the results of the tidal modeling and was $4.7372 \times 10^{-10}$ m$^2$/s, a value equivalent to a hydraulic conductivity of 400 m/d for freshwater at 20°C.

Results and Discussion

Fifteen simulations were completed in which all of the parameters were kept constant except for the value of equivalent hydraulic conductivity, which was varied above and below 400 m/d. Figure 6.14 shows the simulation results for the best-fitting cases. To get the best match between the observed and simulated curves, the section right after the major recharge events was concentrated on. The response of the aquifer to the recharge is much like that produced during a slug test in a well where a volume of water is added instantaneously. The shape of the curve immediately after the recharge pulse depends on the ability of the aquifer to transmit water away from the well. The steeper the curve, the higher the hydraulic conductivity and the faster the mound of water dissipates. The height of the initial rise in water level depends on the specific yield of the aquifer and capillary flow in the unsaturated zone. Neither of these parameters were varied in these simulations.

The simulated transient water-level record for the case where horizontal hydraulic conductivity was equal to 300 m/d seems to provide the best match to the observed data. This value is very close to the value of 400 m/d obtained through the tidal simulations.
Figure 6.14 Observed and simulated water level in Paradise Park Well.

observed water level

recharge = 127 cm/yr

$K = 250 \text{ m/d}$

$K = 500 \text{ m/d}$
Both values seem low compared to other wells outside of the rift and, as was discussed previously, indicate that the Paradise Park Well is located in a region of the aquifer with low hydraulic conductivity. Novak (1995) compared the timing of the rainfall events based on rainfall records from the ERZ and the response of the aquifer, and determined that it took between 8 and 11 hours for the pulse generated by the heavy rainfall events to reach the water table. The distance from the ground surface to the water table is 42.4 m and the vertical infiltration rate is estimated to be about 4.4 m/hr in this area.

6.2.3. Malama Ki Well Simulations

Model Setup

The same mesh design, and boundary and initial conditions discussed previously were used to investigate the transient response of the plume temperature and water level immediately south of the ERZ to recharge changes. The fracture-discharge rate was maintained at the best-fitting value of 5.5 m$^3$/d. Recharge to the model was kept at 88 cm/yr, but scaled by the index factor determined from combined rainfall data from the Malama Ki Farm Station, Pahoa, and PGV records. The Malama Ki Well filtered water-level record shows variations on the order of 5 cm in a month-long time period. Large variations of about 30 cm occur over a six-month time period although the continuous data supporting this may be suspect due to monitoring equipment malfunctions in the well. Temperature variations at the top of the water table were about 4°C during the
five months of continuous monitoring. Long term monitoring shows a total variation of about 5°C over the last 20 years. In contrast, the Paradise Park Well record showed variations in water level of close to 50 cm in one day. Temperature variations at the top of the water table, on the other hand, were only about 0.3°C during the monitoring period. It is apparent that the recharge to the aquifer in the vicinity of the Malama Ki Well occurs much more slowly and causes much smaller water-level responses in the well. The 10-m thick tuff layer from the Puulena Crater eruption probably has a relatively low vertical permeability which slows the infiltration of recharge and smooths out the spiky rainfall input.

An unsaturated-zone vertical infiltration model would be the best way to determine the effects of the unsaturated-zone geologic layering on the rainfall signal. But this type of model would require information about the number and thickness of the tuff layers, and water content versus hydraulic conductivity data for each of the layers. None of this information is available, so a numerical model of this system would be entirely hypothetical. A simpler method of representing the recharge is conceptualized by considering the unsaturated zone as a "filter" which delays and possibly smooths the rainfall signal. To test this idea, the daily rainfall index was smoothed with different moving-average filters of two-day, five-day, and ten-day lengths. The smoothed recharge records were then used as input to the SUTRA energy-transport simulations and the resulting pressure and temperature responses were observed.
Results and Discussion

Four different simulations of transient response to recharge were performed with the four different sets of filtered recharge data. For comparison, the results for the unfiltered recharge and the 10-day filtered recharge are plotted with the rainfall record from the PGV monitoring program (fig. 6.15). Each simulation lasted 23 months from May 1992 until April 1994, a period of time during and after the continuous-monitoring program for the ERZ wells. In the simulation using the unfiltered rainfall index, water levels varied by as much as 20 cm over several days immediately after the heaviest rainfall events. Simulated temperatures at the top of the aquifer changed much less dramatically, (about 2°C), and they occurred over several weeks. As should be expected, higher amounts of recharge caused higher water levels and lower temperatures. In the simulations using the 10-day filtered recharge data, the water level changes were much more subdued. Maximum simulated water-level changes were about 5 cm and these changes occurred over several weeks. Simulated temperatures changed about 2°C, again over several weeks. Both the unfiltered and 10-day filtered recharge input produce almost the same transient temperature results, indicating that temperature results from the model are not very sensitive to rapid changes in recharge.

Comparisons between the simulated pressure and temperature values and the observed values are not particularly good. Some patterns of the observed data are similar to the unfiltered-recharge simulation results and some are like the 10-day filtered results. There are several possible reasons why they differ, most likely due to problems with the data and the conceptual model. The problems involved in collecting the data have been
Figure 6.15 Observed and simulated water level and temperature in Malama Ki Well.
discussed previously in this report and in others (Thomas, 1994 and Novak, 1995). The observed data set is very small and unfortunately exists for a time period where rainfall was not extreme. The observed changes in water level do not correspond very well with the temperature changes, and neither correspond with the rainfall record. For example, the section of observed water level and temperature record around October 1992 is somewhat suspicious. It shows a 10-cm fall and rise in water level and a 2°C drop in temperature at the same time. It is likely that the recording instruments were failing at this time, so neither record can be used with much confidence. Also, the temperature record represents the water temperature at a fixed depth below the ground surface. As the water level changes due to daily tidal and barometric variations and longer term recharge-induced variations, the position of the instrument in the water varies. This leads to variable readings by the instrument because the hot water in the well is constantly moving up and down. There are problems with the model set-up also. Because the model does not take into account unsaturated flow, several processes are ignored. The impact on the recharge as it travels through the thick tuff layers is unknown.

Because there are currently no reliable continuous records of pressure or temperature from the Malama Ki Well, the model cannot be adequately tested in transient simulations. Considering these problems, this current model is probably best suited for simulating long-term changes in temperature due to seasonal recharge variations. In these simulations, average seasonal variations in recharge cause temperature changes of a few degrees at the water-table surface. Both the unfiltered and filtered recharge data sets produce about the same temperature results so it is not critical to determine which works best.
6.2.4. MW-2 Simulations

Model Setup

The same mesh design, and boundary and initial conditions discussed previously were used to investigate the transient response of the plume temperature, water level, and solute concentration in MW-2 to recharge changes. The fracture-discharge rate was maintained at the best-fitting value of 2.3 m$^3$/d. Recharge to the model was kept at 88 cm/yr, but scaled by the index factor determined from the the combined Malama Ki, Pahoa, and PGV rainfall data. It is apparent that the recharge to the aquifer in the vicinity of MW-2 occurs slowly as it does at the Malama Ki Well. It is not known whether or not tuff layers from the Puuolena Crater eruption or other eruptions are penetrated by MW-2. It is very likely that some low-permeability tuff layers occur here due to the well's location in the middle of the rift zone. To take this into account, the smoothed rainfall patterns were also used as recharge input for these simulations.

Results and Discussion

The MW-2 filtered water-level record shows a steady decline in water level of about 35 cm over a four-month time period, April - July, 1993 (fig. 6.16). Temperature variations were about 1°C during the four months of continuous monitoring. The earlier records show water level variations of 70 cm and a total temperature variation of about
Figure 6.16 Observed and simulated water level and temperature in MW-2.
The transient simulation that produced long-term water level variations of about 70 cm had the 10-day filtered recharge input (fig. 6.16). The water level and temperature records from MW-2 collected during the continuous-monitoring program are probably as suspect as the Malama Ki Well records discussed above. The steady decline in water level from April to August 1993 does not correspond with the rainfall record. The highest water level recorded falls at the end of several months of low rainfall. Over the next three months rainfall increased steadily yet the water level continued to decline. There is no obvious correlation between the observed temperature and rainfall either.

The problem with a stationary temperature probe in a moving temperature-stratified water column exists in this well also. An additional complication with MW-2 is caused by the fact that the well is cased throughout the top 6 meters of the aquifer. Although the temperature inside the well should equilibrate to the aquifer temperature through the metal casing, the effects are not readily known. An additional set of water level and temperature data is available from periodic measurements taken as part of the state-required monitoring program (written commun.; DLNR, 1993). These records show a much larger variation in temperature and water level than the continuous records do. This may be due to differences in recharge or due to inconsistencies in the measurement depth of each reading.

The transient simulations do not correspond very well with the observed water-level and temperature trends. But the inconsistencies in the observed data make it very difficult to adequately adjust the model to match the observed data. It is not known whether the observed changes are recharge induced, fracture-discharge induced, or just
due to unreliable measurements. The current model can only be used, at best, for predicting long term trends in well temperature based on recharge. A scenario simulating a drought was performed to show how the temperature at MW-2 might increase independently of changes in the fracture-discharge rate or from geothermal well leaks. To simulate a drought, the recharge to the top of the model was reduced to zero for 90 days. The temperature at the MW-2 location increased by about 3.5°C after 90 days of simulated time.

6.3. Modeling of Regional Energy and Solute Transport

6.3.1. Modeling Approach

The ERZ hydrologic system is very complex and includes fracture and porous media fluid flow, convective and conductive energy flow, and solute transport in the form of a freshwater-saltwater transition zone. An ideal model of the system would incorporate all of these processes in a three-dimensional representation of the system. Unfortunately, several factors make this ideal representation currently impractical to obtain. The most significant factor is a lack of areally distributed data from monitoring wells both in and outside the ERZ. As discussed previously, most of the wells are clustered around the current geothermal development site. Also, efforts at collecting consistent, long-term records of water levels, fluid temperature and fluid concentrations have not been very successful. The resulting model would be based mainly on speculation about the
hydrologic system in the areas lacking sufficient data and there would be no way of validating the results of the simulations.

The second problem with a three-dimensional simulation is the lack of computational power to solve the large numerical simulations in a timely manner. The state-of-the-art three-dimensional numerical models for energy and solute transport are still relatively slow and cumbersome to use even with the most advanced Supercomputing facilities available.

To simplify the study into a more manageable problem, the flow and transport simulations were conducted using two-dimensional vertical cross-sectional meshes oriented both parallel and perpendicular to the ERZ. Both SUTRA and CFEST were used to optimize the modeling effort. SUTRA was used because it is much faster than CFEST and is more robust and stable for the range of parameters needed in this study. The drawback with SUTRA is that it cannot handle energy and solute transport simultaneously. CFEST was used for the final simulations when both processes needed to be considered.

The first step in the regional modeling effort was to simulate the freshwater lens and transition zone in the aquifer using CFEST. The aquifer water temperature was assumed to be constant and only the flow and solute-transport equations were solved. Therefore, changes in fluid density and viscosity were only due to changes in solute concentration. The initial solute concentration of the water in the aquifer was equivalent to seawater chloride concentration and recharge of freshwater was added to the system. The simulation was continued for 30 to 40 years until a freshwater lens and transition zone developed and reached a "steady-state" condition. A steady-state condition represents a condition where fluid solute-concentration at any point remains constant for a significant
length of time such as a year. Observed water-level values and flow gradients from wells both inside and outside the rift were compared to simulated values and a reasonable match was obtained. Values for horizontal longitudinal and horizontal transverse dispersivity length were adjusted to produce a reasonable transition zone based on salinity logs from other Hawai‘i wells. Voss and Souza (in press, 1995) report that these two parameters are the dominant controls on the transition zone width for a solute-transport SUTRA simulation of the Pearl Harbor Aquifer.

The simulated steady-state freshwater lens represents a typical cross-section of a basalt aquifer which includes both a low permeability rift zone and a high permeability coastal aquifer without thermal influences. This simulation was then used to make some predictions about the travel time and flowpath of water in this type of system. This simulation was also used to make some estimates of the amount of freshwater flux in the system.

Originally, after obtaining a steady-state freshwater lens, the plan was to continue using the CFEST model with the addition of energy transport from a water-producing fracture. This step proved to be quite troublesome as some of the drawbacks of the CFEST code became apparent. The model validation tests demonstrated that a SUTRA simulation can be solved for an identical flow problem at least ten times faster than a CFEST simulation. The CFEST element length needed to accurately simulate the rapid changes in temperature and density caused by fracture discharge is less than 10 to 20 m. Such a tight spacing would require a large number of elements in the CFEST mesh and then solution times would increase even more to unreasonable lengths.

Secondly, the CFEST code did not handle energy transport at the top of the aquifer.
in an acceptable manner. It would only provide accurate results when the top boundary of the mesh was set as a specified temperature boundary. This condition did not match the true conditions in the aquifer. Hence, separate SUTRA simulations of fluid flow and energy transport were performed to obtain much of the relevant information about the energy-transport characteristics of the shallow system. These simulations were continued for 40 years of simulated time until a steady-state temperature distribution was obtained.

Finally, CFEST was used to simulate the concentration distribution based on the SUTRA temperature distributions. Subsequent CFEST simulations involved only flow and solute transport with the temperature distribution based on the discharging fracture model remaining constant. This final combination was continued until a new steady-state concentration distribution was obtained.

**Parameter Uncertainty and Sensitivity**

The main problem with modeling ground-water flow in the ERZ is the lack of information regarding the spacing and location of impermeable dikes and highly permeable fractures. Several investigators have attempted to apply stochastic and statistical methods to solute-transport simulation in fractured rock systems (Smith, et al., 1989, Coakley, et al. 1989). Smith et al. modeled transport in discrete fractures based on particle-tracking methods and included the statistical analysis of the results in a larger scale continuum model. The method was effective in simulating patterns of anisotropic dispersion typical of fractured rock systems. But for this method to be successful, it is essential to characterize the hydraulic properties of the individual fractures in the system.
Coakley et al. developed a method for determining the equivalent permeability of a system by statistically analyzing the geometry of void spaces and contact points of rock matrix in a core of quartz monzonite taken from a mine in Sweden. They compared two flow models, Darcy's Law and a Reduced Navier-Stokes equation and found that the former tends to overestimate velocity for tracer studies. There are numerous other studies of fractured systems (for example, Young and Hoerger, 1989, Hestir et al., 1989) which also involved statistical analysis of an extensive set of data of fracture geometry and distribution. The almost complete lack of this type of data for the KERZ renders these investigative techniques useless for this study. No information exists currently about fracture structure, flow properties or locations in the KERZ. Surface geophysical techniques have not been very successful in providing the type of detailed information need to characterize fracture flow. The only effective way of determining this information is through drilling, detailed geologic logging, downhole geophysical logging and detailed surface geophysics. These methods have been somewhat successful in locating the KS-7 fracture beneath the current geothermal development area, but at considerable expense and time. It is not likely that this type of concentrated effort will be focused anywhere else in the KERZ to characterize other sets of fractures.

The same problems are present for characterizing the location and structure of low-permeability dikes in the aquifer. Geologic and hydrologic evidence points to the existence of these features, as the local flow characteristics around any well are controlled by these features and many well measurements reflect only these local conditions, but detailed information useful in constructing a ground-water model is almost nonexistent. Ground water in the rift zone probably occurs in a system of interconnected
“micro-aquifers” bounded by these individual dikes. The dikes are orders of magnitude less permeable than the lavas, but hydraulic connection among an array of dike aquifers allows large areas of a rift zone to be treated as single aquifers in which the controlling permeability is that of the quasi-vertical dikes (Lau and Mink, 1995). For a regional modeling study, this approach is the only practical way to simulate flow through the rift zone.

Mesh Design and Boundary Conditions

For the simulations using CFEST, a 2,698-element rectangular mesh (fig. 6.17) was created for a regional representation of a cross-section of the aquifer. The elements ranged in length from 12.5 m at the coast where ground-water velocities are highest, to 600 m along the most ocean-ward boundary where conditions are constant because of the buffering effect of the ocean. The majority of the elements representing the aquifer were 100 m long. The element depths ranged from less than 10 m at the top of the water table to 90 m at the base of the mesh. The width of the elements remained 1 m (unit width) and the mesh represented a section of the aquifer 650 m deep and 11,500 m long. The right side of the mesh sloped downward to mimic the bathymetry of the ocean floor based on a map produced by Moore (1971). The top boundary was maintained as a constant fluid-flux boundary with recharge ranging from 190 cm/yr at the left side of the mesh to 38 cm/yr at the right side of the mesh. This recharge distribution was consistent with the recharge distribution presented by Takasaki (1993). All recharge had a constant chloride concentration of 10 mg/L based on the findings of a rainfall sampling study performed by
Figure 6.17 Finite-Element mesh used in CFEST regional simulations.
Scholl et al. (written commun., 1992) on Kilauea. The bottom of the mesh was maintained as a no-flow boundary for fluid. The right side of the mesh was maintained as a specified pressure boundary, equivalent to seawater hydrostatic pressure, and a specified concentration boundary, with chloride concentration equal to a seawater values. The initial temperature distribution of the aquifer (fig. 6.18) was based on temperature logs from ERZ wells and from a profile of ocean temperature off of Cape Kumukahi reported by Taguchi et al. (1985).

For the simulations using SUTRA, an 18,981 element rectangular mesh (fig. 6.19) was created for simulating energy transport near the top of the aquifer. Each element was 4.2105 m long, 5.0509 m deep, and 1 m wide (unit width), and the mesh represented a section of the aquifer 80 m deep and 5,000 m long, the approximate distance from the geothermal development area to the coast. The salinity of the water in the top 80 m of the aquifer is very low, so solute-related density and viscosity effects can be ignored. The top boundary was maintained as a constant fluid-flux boundary with recharge of 84 cm/yr at 24°C. The bottom of the mesh was maintained as a no-flow boundary for fluid and a constant-flux boundary for energy of 0.1 J/s (~ 500 mW/m²). The right side of the mesh was maintained as a specified-pressure boundary equivalent to seawater hydrostatic pressure and specified-temperature boundary with reported ocean temperature values. Input to the left side of the mesh, representing fluid flux from up gradient, was 0.152 kg/s, the same as for the fracture simulations. The discharging fracture was simulated as a column of fluid-source nodes located 100 m from the left boundary of the mesh. The injected water in all cases was kept at a temperature of 100°C.
Figure 6.18 Temperature distribution for regional CFEST simulations.
Figure 6.19 Portion of finite-element mesh used in SUTRA regional discharging-fracture simulations.
Input Parameters

Table 6.1 lists the values of fluid and solid matrix properties held constant in all of the regional-flow simulations. The horizontal permeability of the aquifer was held constant at $4.7372 \times 10^{-11}$ m$^2$/s inside the rift, $4.7372 \times 10^{-10}$ m$^2$/s between the rift and flank, and $4.7372 \times 10^{-9}$ m$^2$/s outside the rift (fig. 6.19). These values are equivalent to a hydraulic conductivity of 40, 400, and 4,000 m/d, respectively for freshwater at 20°C. The vertical permeability was kept at $1.4804 \times 10^{-11}$ m$^2$/s (10 m/d) in all three zones. These values were judged to be a good representation of the aquifer hydraulic conductivity distribution based on the tidal-response simulations and subsequent comparison of observed aquifer water levels to simulated water levels from the regional simulations.

The values for solute dispersivity were initially based on reported best-fitting values for aquifer simulations of the Pearl Harbor Aquifer performed by Souza and Voss (1987). They found that the best value for horizontal longitudinal dispersivity was 76 m and the best value for horizontal transverse dispersivity to be about 0.2 m. Subsequent testing with CFEST revealed that smaller values (45 m and 0.08 m) gave the best results for the simulated transition zone in this study. For steady-state conditions, Souza and Voss reported that a transition-zone simulation would be insensitive to both vertical longitudinal and vertical transverse dispersion for a wide range of values tested (Voss and Souza, in press, 1995). Therefore, these values remained the same as the horizontal values for all of the simulations. A search of published data revealed that there are no salinity profiles in the ERZ or any dike impounded aquifer in Hawai'i. Nor was it possible to locate any salinity profiles of the transition zone in high-permeability, young basalt
aquifers like those on the flank of the ERZ. All of the available deep salinity profiles are for aquifers which include a caprock causing the freshwater lens to be thick, unlike the thin lens beneath the south flank of the ERZ. Therefore, it is impossible to try to determine the most appropriate dispersivity values for these aquifers based on observed data.

Values of thermal longitudinal and transverse dispersivity used in the energy-transport simulations were 1.0 m and 0.1 m. Marsily (1986) reports for a single experiment site in Jura, France, that chemical and thermal dispersivity lengths are comparable at the field scale (tens of meters). However, he goes on to state, "..on a larger scale, in a tracer experiment over a longer distance and time, the coefficients of dispersion and conduction reach an asymptotic value, and the dispersivity of the tracer should be around five times stronger than the thermal dispersivity. However, there are no experiments confirming this".

6.3.2. Modeling Results and Predictions

**SUTRA Energy and Solute Transport**

The findings from the simulations of fracture discharge up gradient from MW-2 were used in the regional SUTRA simulations of energy transport. Figure 6.20a shows the "steady-state" temperature distribution in the modeled area after 40 years of simulated time for a fracture-discharge rate of 2.3 m$^3$/d of 100°C water (10.7 kJ/s). The model predicts a high temperature gradient of 0.13°C/m in the first 400 m from the well and a
Figure 6.20 Regional temperature distribution for three fracture scenarios.
much smaller gradient (~ 0.003°C/m) from 400 m down gradient to the coast. Water with a temperature greater than 40°C can be found to a distance of about 800 m from the fracture. The maximum temperature of the water as it reaches the coast is about 30°C. The temperature range of the Pohoiki (or Isaac Hale Park) Spring, located directly down the topographic gradient from MW-2, was 34°C to 35°C in 1991-92 (Sorey and Colvard, 1994). In January 1950, the water had a temperature of 33°C and the springs are said to have gotten a little hotter after the 1955 and 1960 rift zone eruptions (MacDonald, 1973). Without direct evidence though, it is difficult to say much about the timing or location of the source of this water. Although the source of the water discharging at this spring is unknown, the close match between the simulated temperature near the coast and the measured spring temperature suggests that the simulation is a reasonable representation of the system. Figures 6.20b and 6.20c show the simulated "steady-state" temperature distribution for cases b and c where the fracture-discharge rate is raised to 4.6 m³/d (21.4 kJ/s) and 9.2 m³/d (42.8 kJ/s). The temperature gradient in the first 400 m flattens out to 0.11°C/m and 0.07°C/m, respectively, as the core of the plume spreads farther down gradient. For case b, water with a temperature greater than 40°C can be found to a distance of about 2,000 m from the fracture and the maximum temperature of the water as it reaches the coast rises to about 32°C. For case c, water with a temperature greater than 40°C can be found to a distance of about 4,100 m from the fracture and the maximum temperature of the water as it reaches the coast rises to about 36°C.

Figure 6.21a shows a set of three "steady-state" temperature profiles at different distances from the fracture for each of the three different fracture-discharge simulations. At 100 m from the fracture, the approximate location of MW-2, the temperature of the
Figure 6.21  Temperature profiles and transient response for regional energy-transport simulations.
plume rises 27°C, from 61°C to 88°C for the best-fitting case when the simulated fracture-discharge rate is quadrupled. Halfway to the coast, at 2,400 m from the fracture, the temperature rises 17°C, from 31°C to 48°C when the simulated fracture-discharge rate is quadrupled. The maximum change at the coast is only about 5°C. The bottom plot on fig. 6.21 shows the transient temperature response at the three locations. At the MW-2 location, the effects of increasing the fracture-discharge rate are noticeable within a few months and there is not a large time difference between the two simulations. Halfway to the coast, the effects will not be visible for 10 and 7.5 years after the change. And at the coast, it will take 16 and 12 years respectively for the temperature to begin to rise when the fracture-discharge rate is doubled and quadrupled.

**CFEST Solute Transport**

The regional cross-sectional model of flow and solute transport in the rift and flank was performed using three different temperature scenarios; one which maintained the temperature throughout the whole aquifer at 24°C; one which maintained a constant temperature distribution based on a regional flow pattern with no upwelling fractures; and one which included the effects of an upwelling fracture. The first case, with a single temperature everywhere, represents the most simplistic way of modeling a transition zone system and has been used in all previous Hawaii transition zone models. The assumption of a uniform temperature is reasonable because most observed Hawaii ground-water temperatures are fairly uniform and not effected by hydrothermal processes. The second case is more appropriate for the ERZ because of the wide range of
temperatures found in the shallow aquifer (fig. 6.22). The third case is more problematic, as will be discussed later in this section, because it imposes what is essentially a localized flow regime onto a regional scale model of coarser dimension.

The chloride concentration distributions for the first and second case are shown plotted as values relative to the chloride concentration of seawater (fig. 6.23). As expected, the simulated freshwater lens is thickest in the rift zone where permeabilities are lowest and much thinner in the higher permeability flank lava flows. The combination of high permeability and low recharge within 5 km of the coast produces a simulated transition zone with no freshwater lens on the surface. Although the limited observed chloride concentration values in the flank are not quite as high as the simulated values, they all exceed the recommended EPA drinking-water secondary standard of 250 mg/l of chloride. Because there are no records of salinity profiles in the ERZ it was not possible to match the simulated transition zone thickness to observed values. Also, the chloride values that are available for south flank wells are not representative of the basal system because they contain a component of saltwater from the deeper system. Therefore, they are not acceptable as matching points either. Based on the finding of Voss and Souza (in press, 1995), the position of the 50-percent seawater isopleth is a function of the aquifer recharge and hydraulic conductivity and its position is not sensitive to the dispersivity length, so its location is probably the best indicator of the true aquifer conditions.

When the more realistic aquifer temperature distribution is used (second case), the simulated freshwater lens in the rift zone becomes thinner and the position of the 50-percent seawater isopleth is as much as 60 m higher. This difference can be attributed to higher hydraulic conductivity in the aquifer because of the decrease in
Figure 6.22 Temperature distribution for regional CFEST simulation—second case.
Figure 6.23 Chloride concentration for regional CFEST simulations -- first & second case.
temperature-dependent viscosity. Near the coast, where cold water from the ocean is
circulated, viscosities are slightly higher and the transition zone extends slightly deeper
into the aquifer.

The simulated pressures across the top of the mesh were converted to equivalent
head values based on the density at the same node point (fig. 6.24). The plot shows the
importance of using a more realistic temperature distribution in head estimation. For the
first case, the simulated heads south of the rift are one meter or less and rise to about 6.4
meters inside the rift zone. For case 2, the heads outside the rift are about the same as
case 1 but inside the rift, the difference between the two cases is as much as 1.1 m.

There is a problem encountered when comparing simulated heads to observed
heads in the LERZ. The numerical model calculates pressure, not head, because of the
variable density of the aquifer fluid. To get a value of head for a particular element, the
pressure must be divided by the gravitational constant and the fluid density of the element.
The larger the thickness of the element, the more the head result is an averaged value.
The values of head measured in the shallow wells are not for a single point in the aquifer
but a composite over the whole open interval of the well. Most of the shallow wells contain
water that is temperature stratified and therefore density stratified. The observed head
value is therefore not strictly comparable to the simulated head value, and exact matches
would not be expected. The four observed head values shown in figure 6.24 are for
Pahoa Well A, MW-1, MW-2, and Malama Ki Well. The values are average values of
measured head taken from the historical record because none of the wells were ever
measured concurrently on the same day. The values are approximate because the exact
surface elevations at the wells are not known (as discussed previously).
Figure 6.24 Comparison of simulated head results, first & second cases.
CFEST has a subroutine for producing flow vectors based on the steady-state temperature and concentration distribution. Plots of the flow vectors show two patterns, the path of recharge entering at the top of the aquifer and the path of saltwater circulating into the aquifer from the ocean (figs. 6.25 and 6.26). The flow vectors indicate that water entering the aquifer farthest from the ocean takes a flow path only through the freshwater lens as it travels to the ocean. A counter-flow pattern is set up with water flowing into the deep part of the aquifer from the ocean and traveling upward where it flows back toward the ocean. This type of flow pattern is typical for basal lens systems. The flow of freshwater to the ocean carries saltwater from the base of the transition zone with it. To maintain mass balance of saltwater, water must flow into the aquifer from the ocean at depth. The flow-vector plot also indicates zones of travel time in the aquifer shown as the amount of time it would take a "packet" of water to reach the coast from its present position. Most of the water entering the fresh part of the lens as recharge will reach the discharge point at the ocean in less than ten years. Most of the saltwater from the ocean boundary circulates through the system in ten years or less. Travel times are as much as 30 years from the deep part of the rift zone, the area represented by the lower left hand corner of the model. A zone of stagnation occurs in this area where water travels much at a slower rate through the aquifer. Although fractures are not explicitly considered, it is easy to see from the figures how saltwater can migrate into the rift zone where it encounters fractures allowing it to penetrate downward into the deeper geothermal system. Around the area of the rift zone where fractures are expected to occur (4,000-5,000 m from the left side of the model grid), the natural flow gradient is upward which enhances the rise of hydrothermally altered fluids. When the less dense
Figure 6.25 Flow vectors and travel times to ocean for regional CFEST simulation—first case.
Figure 6.26 Flow vectors and travel times to ocean for regional CFEST simulation—second case.
hydrothermal fluid reaches the top of the water table, the natural flow pattern will sweep the water back towards the coast where it discharges.

When a temperature distribution based on the discharging-fracture simulations is used in the regional model (fig. 6.27), the freshwater lens and transition zone and salinity density distribution change (fig. 6.28). The low-density hot water plume creates a mound at about 4,000 to 5,000 m from the left boundary that acts as a local barrier, causing up-gradient heads to increase and the underlying freshwater lens to become much thicker. Simulated ground-water flow patterns are altered as well (fig. 6.29). The upwelling area acts as an effective barrier to saltwater migration from the flanks into the rift zone. Recharge to the top of the aquifer flows much deeper into the aquifer before it migrates upwards and then to the coast. Saltwater does not penetrate as far into the aquifer because of the barrier created by the discharging fracture.

This scenario is useful for investigating the hydrodynamic changes in the system due to the addition of upwelling fluid, but it is not very good for regional flow matching because the model ignores flow perpendicular to the plane of the cross-section. In the aquifer, local mounding would occur over the upwelling fracture as the simulation shows, but because the fracture is not infinite in length, heads would only build up to a certain level before water would flow laterally around the upwelling area and move down gradient. One of the assumptions in using a cross-sectional model is that the cross-section represents a slice of a system that can be considered infinite in length at the scale of study. This assumption is violated when a fracture is added to the simulation. Hence, this simulation is really best suited for demonstrating only the general flow dynamics that the addition of a discharging fracture creates.
Figure 6.27 Temperature distribution for regional CFEST simulation — third case.
Figure 6.28 Chloride concentration for regional CFEST simulation – third case.
Figure 6.29 Flow vectors and travel times to ocean for regional CFEST simulation -- third case.
7. CONCLUSIONS

The primary purpose of this study is to further the understanding of ground water flow and transport in the shallow aquifer of the Kilauea East Rift Zone and adjacent areas. This was accomplished by gathering all of the previously published and newly acquired data and forming a conceptual model of the hydrologic system. The conceptual model was tested by numerical simulations of density-dependent ground-water flow and energy and solute transport using two published finite-element models, SUTRA and CFEST. The conclusions drawn from these simulations can be used as a basis for monitoring the shallow ground-water system and for analyzing potential future effects of geothermal development.

7.1. Conceptual Model

There are two principal components of the Kilauea East Rift Zone ground-water system: the geologic framework and the hydrodynamic flow system in this framework. The regional geologic framework is fairly well understood based on years of investigation involving field mapping, well drilling and geophysical surveys, but the features such as intrusive dikes, fracture zones, and vertical faulting, that control ground-water flow on a local scale are poorly delineated. The presence of low-permeability dikes causes head buildup in the rift zone as water is impounded behind the dikes and compartmentalized into "micro-aquifers". Fractures and faults bisect the dike zones and allow cool freshwater
to migrate downward to a deeper geothermal system where it is heated by energy released from cooling magma. Different sets of fractures act as conduits for the now heated water and steam to rise back into the shallow aquifer. The heated water floats on the shallow water table due to its lower density and migrates down gradient with the regional flow pattern. At least one such deep-system discharge area exists around Puu Honuaula, which lies next to several vents of the 1955 lava flow, and another exists around Puulena Crater, a phreato-magmatic eruption feature. The conduits for steam and hot water discharge around Puu Honuaula are probably fractures corresponding to the sides of a graben which is still being downthrown. Comparisons of temperature profiles from 20 years ago with recent profiles show the temperature profiles have not changed over the last 20 years and suggest that the system appears to be at a steady-state condition. The existence of other discharge areas is likely but more evidence is needed to locate the fractures associated with upwelling at these areas. The decreasing frequency of intrusive dikes and large-scale fracture zones in the LERZ allows water in the rift to flow to the south into the higher-permeability flank basal ground-water system. Here, the water has sufficient heat to remain above ambient ground-water temperatures in south flank wells and in basal springs at the coast.

The continuous monitoring of several shallow wells in the LERZ provided information about the tidal and atmospheric influences on the shallow water table. The shallow water table both inside and outside the rift zone fluctuates due to: 1) the effects of daily tidal variations in the ocean, 2) barometric pressure changes in the atmosphere, and 3) recharge infiltrating through the unsaturated zone. The tidal variations decrease with increasing distance from the coastline and as the frequency of dikes increases. The
rift-zone lavas in the first 5-10 km up rift from Cape Kumukahi transmit the tidal signal just as efficiently as the flank lavas, suggesting a lower frequency of dikes or higher frequency of fractures in this region of the rift zone. The tidal signal is unexpectedly low in the Paradise Park Well suggesting either poor communication between the well annulus and the aquifer or a region of low permeability between the well and the coast. The tidal signal and the barometric pressure signal in any well is large enough to mask the changes in water level due to recharge. These effects must be removed before recharge-induced water-level changes can be determined. An effective method for removing these effects based on frequency analysis and correlation with ocean levels has been demonstrated.

The infiltration of recharge from the surface to the water table, a distance of about 40 m, takes only about 10 hrs in the vicinity of the Paradise Park Well. The low permeability of the aquifer in this area allowed a rapid rise in water level of as much as 0.5 m after a particularly heavy recharge event. Thick layers of tuff from the Puulena eruption in the vicinity of the Malama Ki Well and possibly MW-2 slow down the infiltration rate of recharge. No water level changes can be directly correlated with rainfall events but both the water level records and rainfall records are incomplete. Because the permeability of the aquifer in the vicinity of the Malama Ki Well is relatively high, water-level changes due to extreme recharge events are very subdued.

Green Lake, the only surface water body in the LERZ, receives recharge from direct rainfall and through infiltration in the Kapoho Crater tuff cone. The lake water is perched on a low permeability tuff layer and is not directly connected to the basal aquifer. Water is lost from the lake through evaporation and presumably through infiltration through the tuff layer into the basal system. The Kapoho Shaft Well also collects water
that is perched on the tuff layer and does not appear to be connected to the regional aquifer. Both the lake and the well show low tidal responses even though they are only about 2 km from the coastline. This constitutes more evidence that they are separated from the basal system.

7.2. Numerical Simulations

Simulations designed to verify results from CFEST were successful for one-dimensional solute and energy transport. CFEST was also shown to accurately match the results from an analytical solution derived to estimate the thickness of a freshwater lens based on sharp-interface approximations. When comparisons of identical SUTRA and CFEST simulations were made, it was shown that, as the simulations became more complicated and involved temperature-dependent fluid viscosity and fluid density, CFEST results lost accuracy. The errors were significant enough that CFEST was not used for additional energy transport simulations.

Numerical simulation of the propagation of the ocean tidal signal through the aquifer provided information about the hydraulic conductivity and specific yield of the rift zone and flank lavas. In order to simulate the observed tidal responses in monitoring wells, the hydraulic conductivity (assuming fresh, 20°C water) of the aquifer south of the rift zone must be between 1,000 and 6,000 m/d. This same range of values also applies for the LERZ between the Kapoho Airstrip Well and the coast, a distance of about 5 km. To reproduce the low tidal response inside the rift zone, the hydraulic conductivity farther uplift must be at least two orders of magnitude lower than hydraulic conductivity in the
coastal portion of the rift zone and outside the rift. To mimic the tidal lag in simulations, the specific yield or "effective porosity" must be between one and three percent. Simulations of recharge events in the north flank aquifer (Paradise Park Well) suggest that the hydraulic conductivity there is between 200 and 400 m/d.

Detailed small-scale numerical simulations of the addition of hot water to the shallow aquifer from postulated fractures were completed for three of the wells with good measured temperature profiles: GTW-3, MW-2, and the Malama Ki Well. To simulate the temperature profile measured in GTW-3, the simulated fracture had to be discharging about 53 kJ/s (26 MW) of fluid (11.7 m³/d of 100°C water) a distance of about 170 up gradient from the well location. Similar simulations to match the temperature profile in MW-2 suggest that the well is about 105 m down gradient from a fracture discharging 11 kJ/s (~ 5 MW) of fluid (2.3 m³/d of 100°C water) into the shallow aquifer. Simulations to recreate the Malama Ki Well temperature profile suggest that this well is located about 700 m down gradient from a fracture discharging about 26 kJ/s (12.8 MW) of fluid (5.5 m³/d of 100°C water). Puulena Crater, containing active steam vents, is about 700 m up gradient from the Malama Ki Well and is assumed to be the source of hot water measured in this well.

Simulations of the transient response to changes in fracture-discharge rates provide valuable information about the utility of the existing wells as monitoring points. Changes in the volume of hot fluid being added to the shallow aquifer, either through geothermal-development induced changes in the fracture-discharge rate or through leaks in the geothermal production or injection wells, will produce changes in the temperature measured in wells located down gradient. Up gradient from MW-2, the most important
monitoring point, a reduction in discharge from a fracture due to geothermal production would cause a change in temperature at the well. In simulations, when the fracture discharge was reduced to zero, it took about 130 days for the temperature at the MW-2 location to drop at least 3°C from the initial temperature. Smaller reductions required even longer time for significant changes to be felt. Simulations designed to predict the effects of a leaky geothermal production or injection well suggest that significant changes in temperature at the well location would be felt between ten and forty days after the onset of a major leak.

To test the sensitivity of SUTRA simulations, seven parameters were varied by as much as an order of magnitude and the results were compared. Transient energy transport simulations are most sensitive to changes in both longitudinal and transverse dispersivity length and are not very sensitive to changes in rock density, porosity or anisotropy ratio. For the most realistic representation, dispersivity length should be very low or even zero, based on the Peclet number for energy transport. In this range, the energy-transport model is not very sensitive to changes. Transient solute transport simulations are most sensitive to the value of porosity and less sensitive to changes in dispersivity length. The relative insensitivity of the steady-state simulations to changes in solid matrix and aquifer parameters suggest that the fracture-discharge simulations are fairly robust, and are mainly dependent on fracture-discharge rate and distance to the observation well.

Regional energy-transport simulations show that at the coast, the temperature of the warm springs would rise about 5°C for the scenario in which the rate of energy discharge into the shallow aquifer up gradient of MW-2 was increased fourfold. The
simulations suggest that these effects would not be felt at the coast for about 12 years. Lesser changes in the fracture-discharge rate show correspondingly lesser effects at the coast after greater lengths of time. These simulations are for "worst case" scenarios where fluid leaks into the shallow aquifer at high rates for the entire length of the simulation. Realistically, a leak would be noticed immediately at the geothermal well location and discharge to the aquifer would only last a few hours or days. Therefore, the modeled scenario is for the worst case and the impact on the aquifer would be much less than the simulation predicts.

Attempts to match observed temporal water-level and temperature changes with simulated results were not entirely successful. The simulations are useful for showing seasonal changes based on recharge patterns, but not for short-term changes. A simple drought scenario with no rainfall was simulated for a period of three months and the temperature at the MW-2 location increased by about 3°C. The lack of reliable or continuous water-level and/or temperature records makes it nearly impossible to say whether or not the model is valid. Temporal monitoring records that are much more reliable are needed before the transient models can be properly "proven" to work with detailed rainfall data.

Regional modeling of solute transport using CFEST was used to investigate the effects of using different temperature distributions on the shape of the freshwater lens. Three different temperature distributions were used; a uniform distribution of 24°C, a more realistic regional distribution, and a distribution created by a discharging fracture. The first two cases produce similarly shaped freshwater lenses, but the lens produced by the second case is thinner because of lower fluid viscosity in the aquifer. These results are
significant for future ground-water availability modeling studies where ignoring temperature effects may lead to overestimation of the resource. The third case produces a freshwater lens that is much thicker because the discharging fracture acts as a barrier to keep saltwater out of the rift and allow freshwater heads to build up inside the rift. The simulated travel times for water to reach the ocean are less than 10 years for most of the recharge entering the aquifer.

7.3. Suggestions For Improvement and Future Work

There are several improvements that can be made to the East Rift Zone conceptual and numerical ground-water models. The biggest improvement would be to create a three-dimensional numerical model of the system to simulate energy and solute transport. This would provide a better understanding of the complicated flow dynamics associated with upwelling hydrothermal fluid. This would also allow simulations that could demonstrate the relationship between ground-water flow that occurs parallel to the rift zone and flow that occurs perpendicularly. In reality, the current three-dimensional, energy and solute-transport models are still too slow to be efficiently used for the conditions found in the LERZ, namely a hydrothermal system interacting with a freshwater/saltwater transition zone. The current version of CFEST has proven to be inadequate because of computation speed and inaccurate results using LERZ aquifer and ground-water parameters.

One problem with modeling regional flow patterns which may be quite difficult to overcome is that the discharge of hot fluid to the shallow aquifer occurs in single fractures
and creates a very localized temperature perturbation in the water table. To properly simulate this temperature distribution, a numerical mesh must have a small element size, on the order of 5-10 m. The regional model must cover 5-10 km to incorporate all of the important boundary conditions. Satisfying both requirements with a single mesh would mean creating a huge grid of elements. This would require considerable time to create the input files, and computation time would be tremendous. Current numerical models can theoretically handle this size of problem, but in practical terms, they are not really capable of producing results which can be applied in a timely manner.

If a better, faster model becomes available, there will still be much difficulty in modeling the system because of the scarcity of reliable data. Continuous long-term monitoring of the water level, temperature, and salinity of the existing shallow wells is needed to better delineate the effects of recharge and hydrothermal discharge. Several years of data which includes some extreme recharge events would be ideal. Additional wells between SOH-4 and the geothermal production area would be useful to provide data on this area of the aquifer where the flow gradient is very steep. Additional aquifer testing would be useful for confirming the estimates of permeability based on tidal response. There have been no useful aquifer drawdown tests performed on any of the shallow LERZ wells. The most promising existing location for a pumping test would be at the MW-1 and MW-3 area since they are only about 100 m apart. A useful experiment would be to withdraw water from one of the wells while monitoring the temperature profile in the other well for any changes. This information might show what zones of the aquifer are controlling flow.

Salinity profiles inside and outside the rift zone would also be useful for comparing
to simulated results. The nature of the transition zone inside the rift zone has not been
due to the lack of deep salinity profiles and it is not clear if
ground-water heads in the rift zone are an accurate indicator of the depth of the
50-percent seawater isopleth. Better delineation of fractures and dikes within the rift
would also be helpful because, as this study has shown, regional flow patterns are a factor
of many localized features. A single fracture discharging hot fluid can produce
temperature effects which are felt several kilometers from the source. For the flow
dynamics to be understood completely, all of the local features must be identified.

As part of the overall study of the KERZ geothermal system, the State contracted
with a private geothermal consultant to produce a modeling study of the geothermal
reservoir. Initially, this study was designed to be a parallel effort with that study, and the
findings from each study were to mesh into an overall conceptual model of the flow
system. Unfortunately, proprietary data considerations between the consultant, the State
and the geothermal developers did not allow this to happen. There was very little
interaction between investigators for the two studies, and currently, the results of the
private consultant’s study are unavailable. Therefore, one of the primary sources of
information initially intended to be a basic component of this study was not available.

If the results of the independent modeling study of the deep geothermal reservoir
become available, they could be compared to the findings from this study. Some of the
questions in the conceptual model that could be answered by an integration of the two
studies include:

1) how much fresh water recharge is reaching the deeper system?
2) how much saltwater is entering the deeper system from south of the rift zone?

3) is the recharge to the deeper system localized in convection cells or does it reach the deep aquifer some distance up rift and migrate down rift where it ascends?

4) how much energy is transmitted from the deeper system to the shallow aquifer, and in what state, water or steam?

5) how will geothermal production affect the amount of energy transmitted to the shallow system?

6) how long will production continue at the current facility and how might that change the relationship between the deep and shallow systems?
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210


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212


