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RAM2 MODELING AND THE DETERMINATION OF SUSTAINABLE YIELDS OF HAWAII BASAL AQUIFERS

Clark C.K. Liu

October 2007

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[12] ABSTRACT (PURPOSE, METHOD, RESULTS, CONCLUSIONS)

RAM2 is a simple groundwater flow and transport model developed in a recent research study on the sustainable yield of Pearl Harbor aquifer. This research continues the effort by improving upon the modeling procedures, especially regarding parameter identification. In this research effort, RAM2 is applied in a re-evaluation of the sustainable yield of three principal Hawaii basal aquifers, including Honolulu aquifer in southern Oahu, Iao aquifer in western Maui, and Kualapuu aquifer in central Molokai. In the case of the Honolulu aquifer, the respective sustainable yields of its four separate basal-water bodies (i.e., Moanalua, Kalihi, Beretania, and Kaimuki) are re-evaluated individually. RAM2 modeling consists of a flow submodel and a salinity transport submodel. The transport simulation involves calculating the minimum equilibrium hydraulic head of a basal aquifer, which is the hydraulic head required to prevent the salinity of the water pumped out of the aquifer from being higher than the acceptable level. The calculated minimum equilibrium hydraulic head is then introduced to a parabolic relationship, derived by flow simulation, to calculate the aquifer's sustainable yield. Effective dispersion coefficient is the most important transport parameter for RAM2 modeling. A field tracer method for estimating effective dispersion coefficients of a basal aquifer by using salinity as a natural conservative tracer was developed by this study. Using this method, the effective dispersion coefficient for the basal aquifer at each monitoring well can be estimated by a curve-fitting process that compares the salinity profile observed at a deep monitoring well with calculated profiles. The application of this method requires prior knowledge of the time it takes for a hypothetical water particle to move from the upper boundary of a basal aquifer to a monitoring well, i.e., the time-of-water travel. The quasi-three-dimensional model SHARP, developed by U.S. Geological Survey, was used to simulate groundwater flow in a basal aquifer to calculate the travel time. Re-evaluation of sustainable yield by RAM2 modeling provided the following estimates: 19.0 mgd for the Iao aquifer on Maui, 5.0 mgd for the Kualapuu aquifer on Molokai, and 46.5 mgd for the Honolulu aquifer on Oahu. The approximate sustainable yields of the individual basal-water bodies in the Honolulu aquifer are 17.3 mgd for Moanalua, 8.7 mgd for Kalihi, 14.0 mgd for Beretania, and 6.5 mgd for Kaimuki.

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Project Completion Report
for
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Sustainable Yields of Hawaii Basal Aquifers"

Project Period: 30 June 2005–31 December 2006

Principal Investigator: Clark C.K. Liu

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Effective dispersion coefficient is the most important transport parameter for RAM2 modeling. A field tracer method for estimating effective dispersion coefficients of a basal aquifer by using salinity as a natural conservative tracer was developed by this study. Using this method, the effective dispersion coefficient for the basal aquifer at each monitoring well can be estimated by a curve-fitting process that compares the salinity profile observed at a deep monitoring well with calculated profiles. The application of this method requires prior knowledge of the time it takes for a hypothetical water particle to move from the upper boundary of a basal aquifer to a monitoring well, i.e., the time-of-water travel. The quasi-three-dimensional model SHARP, developed by U.S. Geological Survey, was used to simulate groundwater flow in a basal aquifer to calculate the travel time.

Re-evaluation of sustainable yield by RAM2 modeling provided the following estimates: 19.0 mgd for the Iao aquifer on Maui, 5.0 mgd for the Kualapuu aquifer on Molokai, and 46.5 mgd for the Honolulu aquifer on Oahu. The approximate sustainable yields of the individual basal-water bodies in the Honolulu aquifer are 17.3 mgd for Moanalua, 8.7 mgd for Kalihi, 14.0 mgd for Beretania, and 6.5 mgd for Kaimuki.

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NOMENCLATURE

σ = aquifer thickness; for a basal aquifer with a sharp saltwater–freshwater interface, $\sigma = 41h$

ω = constant width of the aquifer

η = typical deep pumping well depth plus the upconing height, or $\eta = h_{\text{well}} + h_{\text{upconing}}$

ρ = water density

τ = mean hydraulic residence time, which is the time a typical water particle stays in the aquifer (can be defined as $\tau = V/Q$)

ζ = vertical distance from the center of the transition zone to its upper limit, which is defined as the location in a transition zone where water salinity is 2% of saltwater salinity

A = cross-sectional area of the aquifer

C = salinity (‰)

C_0 = saltwater salinity (‰)

D = rate of net draft (pumping minus irrigation return flow)

D_{sus} = estimated sustainable yield

D_x, D_y, and D_z = dispersion coefficients in x, y, and z directions, respectively

h = hydraulic head

h_0 = initial hydraulic head of the aquifer

h_e = minimum equilibrium hydraulic head required to protect sustainable yield

h_{upconing} = upconing height

h_{well} = typical deep pumping well depth

I = rate of natural rainfall recharge

k = hydraulic conductivity

L = leakage rate

n = ratio of net draft and natural recharge, $n = D/I$

Q = rate of outflow from the aquifer (ft^3/s)

q = specific water flux (flow per unit aquifer width)

S = storage coefficient

t = time

t_R = time-of-water travel required for a water particle to travel from the most upstream boundary of a basal aquifer to any deep monitoring well location x at constant flow velocity u (can be defined as t_R = x/u)

u, v, and w = groundwater flow velocities in x, y, and z directions, respectively

V = total water volume of the aquifer (ft^3)

V_0 = total water volume of the aquifer before development (ft^3)

W = source and sink terms (D, I, and L)

x = longitudinal travel distance

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1. INTRODUCTION

1.1. Overview of Groundwater Hydrology in the Hawaiian Islands

The Hawaiian islands originated from lava effusing from deep fissures in the Earth's crust, forming volcanic rocks that are either extrusive or intrusive. Extrusive rocks are derived from magma that is poured out or ejected at the Earth's surface. Intrusive rocks are derived from magma forced into older rocks deep within the Earth's crust, where they slowly solidify.

In Hawaii, extrusive rocks are usually comprised of a caldera and several narrow rift zones that radiate from it. A caldera is a large crater formed by volcanic explosion. A rift zone is usually less than three miles in width and tens of miles in length. The most widespread intrusive rocks in Hawaii are dikes that formed when intruding molten magma solidified in conduits within the volcano's rift zone. These conduits may feed eruptions on the surface or may stay beneath the surface. Dikes are nearly vertical and behave as a low-permeability vertical barrier to groundwater flow in otherwise high-permeability lava rocks.

Fresh groundwater in Hawaii may occur in dike-impounded aquifers and in basal aquifers. Dike-impounded or high-level aquifers occur between dikes in the interior portions of islands, where most of the natural recharge takes place (Figure 1). Freshwater stored in dike-impounded aquifers may leak to basal aquifers by subterranean flow through dike fractures. It accounts for a significant part of the recharge that the downgradient basal aquifers receive.

Basal aquifers are the principal water supply sources for Hawaii. Groundwater in these aquifers occurs in dike-free lava flow formations in the flanks of the volcanoes (Figure 1). The water level of a basal aquifer, which is no more than about 40 ft above mean sea level, is much lower than that of a high-level aquifer. The water level of a high-level aquifer can be as much as 3,300 ft above mean sea level on Maui and Hawaii and as much as 1,600 ft above mean sea level on Oahu (Lau and Mink, 2006).

Perched groundwater bodies are embodied in areally limited and poorly permeable strata within the unsaturated zone. Perched aquifers are generally small and isolated; hence, they are not reliable as a source of water supply, except to provide for a few households.

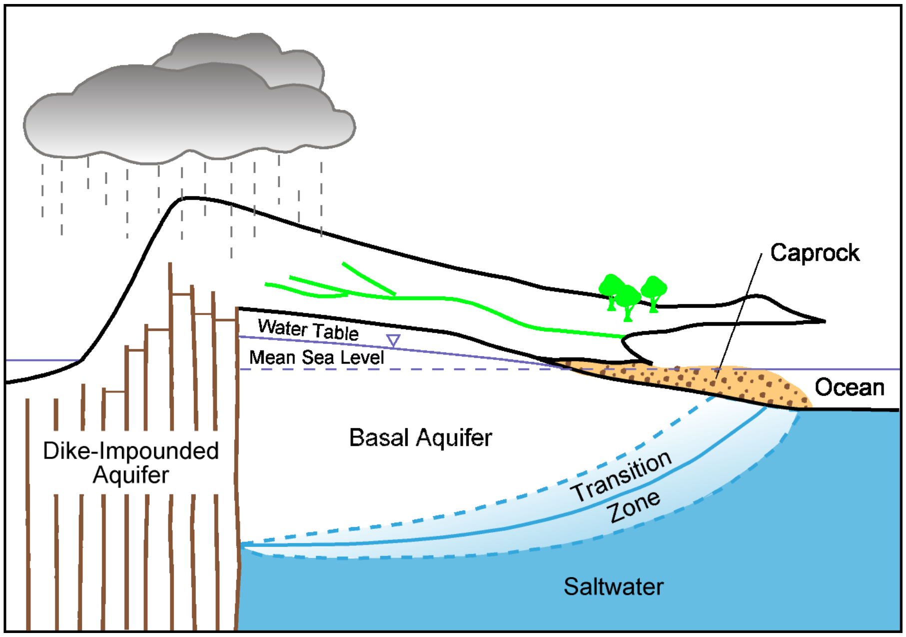


Figure 1. Hydrogeological features of a typical Hawaii basal aquifer

In a homogeneous basal water body in hydrostatic equilibrium, freshwater and the underlying saltwater are separated by a relatively sharp interface. According to the classic Ghyben–Herzberg relationship, the sharp interface of a basal water body is found below mean sea level at a depth of about 40 times the hydraulic head. However, under natural conditions, there exists a transition zone in which salinity decreases gradually from saltwater below to freshwater on top. The thickness of a transition zone depends on transport processes of salt advection and dispersion, which are caused by tidal fluctuation, atmospheric pressure variation, and groundwater recharge. For a preliminary investigation of a basal aquifer resource, the sharp interface assumption is acceptable and the 50% salinity contour in the transition zone can be taken as the approximate location of the sharp interface. However, for the formulation of a detailed coastal groundwater development plan, expansion of the transition zone caused by saltwater intrusion must be taken into account.

1.2. Overview of Numerical Groundwater Modeling in Hawaii

In Hawaii, the first numerical model was developed by GE-Tempo. It was applied in the simulation of long-term head variation of the Palolo aquifer on Oahu (Meyers et al., 1974). Later, the head variation of the Pearl Harbor aquifer on Oahu in response to pumping stress was simulated using a two-dimensional flow model (Liu et al., 1983). Both of these models were formulated by assuming a sharp interface exists between freshwater and the underlying stationary saltwater. These models also assume that there is no vertical freshwater movement (known as the Dupuit assumption). Results of modeling analysis using the two-dimensional flow model indicate that the Pearl Harbor aquifer's response to pumping stress involves the simultaneous adjustment of leakage, storage, and hydraulic head.

In 1978, the U.S. Geological Survey (USGS) started the Regional Aquifer System Analysis (RASA) program in response to a congressional mandate to develop quantitative appraisals of major groundwater systems in the United States. Numerical modeling is a principal component of the RASA program. The Hawaii study was started in 1982 as one of 25 studies under the RASA program (Nichols et al., 1996). The regional aquifer system in Hawaii, consisting of the entire island of Oahu, is divided into seven major groundwater areas.

During the last 20 years, four numerical models developed by the USGS have been applied in groundwater investigations of various Hawaii aquifers. Two of these models, SHARP and AQUIFEM, simulate only groundwater flow. The other two models, MOC and SUTRA, simulate both flow and chemical transport.

SHARP is a quasi-three-dimensional, numerical finite-difference model. It simulates the movement of freshwater separated from the underlying saltwater by a sharp interface (Essaid, 1990). SHARP was used in a USGS study of the effects of additional pumping on groundwater flow in central Oahu (Oki, 1998). In this study, SHARP was used to calculate time-of-water travel from upper aquifer boundaries to individual deep monitoring wells in the Iao aquifer on Maui and the Kualapuu aquifer on Molokai.

AQUIFEM is a two-dimensional finite-element groundwater model. It simulates the movement of freshwater and the underlying saltwater, separated by a sharp interface. Because the movement and head distribution of both freshwater and saltwater are important for an accurate determination of the location of the sharp interface, AQUIFEM may provide satisfactory flow simulation, especially for the short-term response of a basal aquifer to pumping stress. AQUIFEM was first applied in Hawaii to study the change of hydraulic head in the Waialae aquifer system on Oahu in response to increased pumping. Subsequently, it was also applied in modeling studies of the Ewa–Kunia aquifer on Oahu (Souza and Meyer, 1995) and the Kualapuu aquifer on Molokai (Oki, 2000).

MOC was originally developed as a numerical two-dimensional groundwater flow and transport model. The three-dimensional version is now available. The model's governing equations are solved using a combination of the finite-difference method and the method of characteristics to reduce the problem of numerical dispersion (Konikow and Bredehoeft, 1978). MOC is less flexible than AQUIFEM in solving problems involving irregular boundaries. MOC was used by Orr and Lau (1987) to model the transport of dibromochloropropane pesticide residue in the Pearl Harbor aquifer.

SUTRA was originally developed as a two-dimensional finite-element groundwater flow and solute transport model (Voss, 1984). Two-dimensional SUTRA was applied to two RASA program studies in Hawaii. In a study of the Pearl Harbor aquifer, SUTRA was used to investigate the variations of salinity distribution due to pumping stress, especially salinity changes in the vicinity of pumping wells (Souza and Voss, 1987). Later, for the Beretania aquifer on Oahu, SUTRA was used to study the expansion of the transition zone due to increased pumping (Liu et al., 1991a).

Recently, a flow and salinity transport simulation of the Pearl Harbor aquifer using a three-dimensional version of SUTRA was conducted by the Hawaii District of the USGS Water Resources Division (Gingerich and Voss, 2005).

Also recently, three-dimensional groundwater flow and salinity transport modeling of the Honolulu aquifer in southern Oahu was conducted by consultants under contract to the Honolulu Board of Water Supply (Todd Engineers, 2005). The model, a coupled flow and salinity transport simulation, is based on the computer code FEFLOW (WASY, 2002). The FEFLOW model was used to investigate the vertical movement of the transition zone of Honolulu aquifer in response to varying pumping rates. It is noted that the FEFLOW model was not used to re-evaluate the existing sustainable yield estimation of the Honolulu aquifer. Instead, the existing sustainable yield of the Honolulu aquifer—as estimated previously by a simple analytical model called RAM and included in the last Hawaii Water Resource Protection Plan (CWRM, 1990)—was adopted by Todd Engineers.

Although the numerical groundwater modeling efforts mentioned above have enhanced our understanding of Hawaii basal aquifers, most of them are too complicated for users to comprehend. Added to this problem is the fact that adequate parameter identification, calibration, and verification of these models require extensive field data that are not currently available. It is noted that successful model calibrations were claimed with FEFLOW modeling of the Honolulu aquifer (Todd Engineers, 2005) and with three-dimensional SUTRA modeling of the Pearl Harbor aquifer (Gingerich and Voss, 2005), despite the fact that values of flow and transport parameters used in these studies are quite different.

Table 1 shows the different values of effective porosity and dispersivity used in the two modeling analyses. For an unconfined aquifer (such as a basal aquifer in dike-free basalt), the effective porosity or specific yield equals the storage coefficient, which is defined as the volume of water an aquifer releases from or takes into storage per unit area of the aquifer per unit change in hydraulic head. The value of effective porosity used for FEFLOW modeling of the Honolulu aquifer suggests that the aquifer can yield 0.20 ft^3 of freshwater per square foot of surface as the hydraulic head declines by 1 ft, while the value of effective porosity used for SUTRA modeling of the Pearl Harbor aquifer indicates that the aquifer can yield only 0.04 ft^3 of freshwater per square foot of surface as the hydraulic head declines by 1 ft. This is inconsistent with the findings of past studies that the Honolulu aquifer and the Pearl Harbor aquifer, collectively called the southern Oahu aquifer, are hydrogeologically similar.

Table 1. Values of Aquifer Parameters Used in Two Recent Three-Dimensional Modeling Analyses

Parameter	SUTRA Modeling of Pearl Harbor Aquifer (Gingerich and Voss, 2005)	FEFLOW Modeling of Honolulu Aquifer (Todd Engineers, 2005)
Effective Porosity		
Dike-free basalt	0.04	0.20
Caprock	0.04	0.10
Hydraulic Conductivity		
Dike-free basalt (horizontal)	1,500 ft/d	1,500 ft/d
Caprock	1.5 ft/d	1.0 ft/d
Longitudinal Dispersivity		
Dike-free basalt (horizontal)	820 ft	164 ft
Dike-free basalt (vertical)	32.8 ft	—
Caprock	820 ft	164 ft
Transverse Dispersivity		
Dike-free basalt (horizontal)	0.82 ft	16.4 ft
Dike-free basalt (vertical)	0.82 ft	—
Caprock	0.82 ft	16.4 ft

The longitudinal dispersivity used by FEFLOW is 164 ft and that used by SUTRA is 820 ft. The transverse dispersivity used by FEFLOW is 16.4 ft and that used by SUTRA is 0.82 ft. The use of these different dispersivity values would result in a different simulated response of the transition zone for the respective basal aquifers to the pumping stress.

Until consistent model parameters are identified and used, comprehensive numerical models cannot be used as viable analytical tools in groundwater resource management.

1.3. Overview of RAM2 Modeling

A simple robust analytical model (RAM) was developed by Mink (1980, 1981) for determination of the sustainable yield of Hawaii basal aquifers. RAM, which was derived based on many simplifying assumptions, is able to calculate temporal variations of the hydraulic head in a basal aquifer in response to pumping stress. As the most popular groundwater management tool in Hawaii, RAM has been used to estimate the sustainable yield of the Pearl Harbor aquifer and many other basal aquifers in the state (CWRM, 1990).

A basal aquifer reaches a hydraulic steady state when the recharge it receives equals its leakage plus pumping, or forced draft. As indicated by RAM, a parabolic relationship of hydraulic head and draft rate exists when a basal aquifer is at a steady state. Figure 2 shows a plot of head vs. draft in terms of dimensionless variables. The ordinate is a dimensionless

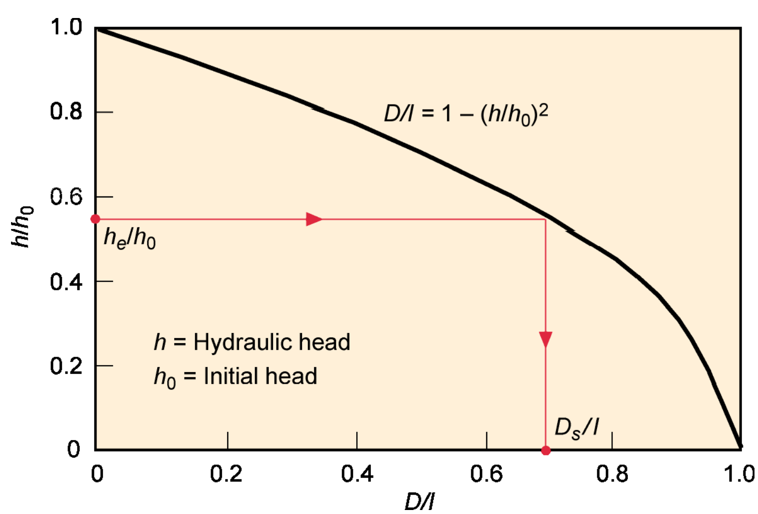


Figure 2. Basal aquifer head–draft curve derived by RAM to determine sustainable yield

variable of head, or h/h_0, where h is the average hydraulic head of a basal aquifer and h_0 is its initial head. The abscissa is a dimensionless variable of the draft rate, or D/I, where D is the draft rate and I is the rate of natural recharge. Note that the initial head of a basal aquifer exists under natural conditions, when rates of recharge and leakage are in balance. Thus, the initial head represents the overall hydraulic properties of a basal aquifer.

According to RAM, the sustainable yield of a basal aquifer relates directly to its minimum equilibrium head. Mink (1980) stated that "the clearest expression of sustainable yield is that of allowable net draft for a selected (minimum) equilibrium head." After the value of equilibrium head (h_e) and thus h_e/h_0 is selected, this value is inserted into the ordinates in Figure 2 to obtain the corresponding dimensionless variable of draft, or D_s/I. Multiplying this value by the known groundwater recharge rate gives the sustainable yield.

The minimum equilibrium head of a basal aquifer cannot be determined analytically by solving the governing flow equation of RAM. Therefore, estimating the sustainable yield of Hawaii basal aquifers by RAM is based on the empirical relationships for selecting minimum equilibrium head, as suggested by the Hawaii Commission on Water Resource Management (CWRM, 1990) (Table 2).

Table 2. Relationships Between Initial Head and Minimum Equilibrium Head of Hawaii Basal Aquifers, Established by CWRM (1990)

Range of Initial Head, h_0 (ft)	Ratio of Minimum Equilibrium Head and Initial Head, h_e/h_0
4–10	0.75
11–15	0.70
16–20	0.65
21–25	0.60
> 26	0.50

Mink (1980) noticed the empirical nature of RAM modeling, which is associated with the selection of the minimum equilibrium head. He suggested that an objective determination of the equilibrium head can be achieved by combining the groundwater flow and salinity transport equations.

In a recent study sponsored by CWRM, a modified RAM, or RAM2, was developed (Liu, 2006). RAM2 consists of two submodels. The flow submodel takes the form of RAM as derived by Mink (1980). The transport submodel simulates the expansion of the transition zone as the head declines. The modeling analysis by the transport submodel of RAM2 calculates analytically the minimum equilibrium head, which is the hydraulic head of a basal aquifer required to prevent the salinity of water pumped out from becoming higher than the acceptable level.

1.4. Research Approach

In this study, RAM2 is used to determine the sustainable yield of Hawaii basal aquifers. Major research effort is made to estimate RAM2 parameters, namely, dispersion coefficient and mean hydraulic residence time. These parameters characterize flow and transport processes in basal aquifers.

A parameter identification method, which calculates the dispersion coefficient of a basal aquifer based on deep monitoring well data, was derived by the recently completed CWRM project (Liu, 2006). The application of this method requires prior knowledge of time-of-water travel between the upper boundary of the basal aquifer and the monitoring well. During this research, it was found that an in-depth study of groundwater flow in large basal aquifers, such as the Iao aquifer on Maui and Kualapuu aquifer on Molokai, would be necessary for accurate time-of-water travel determination. It was also found that an in-depth study of groundwater flow can be achieved by flow simulation based on the sharp interface assumption. The quasi-three-dimensional model SHARP, which assumes the existence of a sharp interface between the freshwater and saltwater zones of a basal aquifer, was selected for flow simulation. The adopted research approach, which includes flow simulation as a prelude to RAM2 parameter identification, is illustrated in Figure 3.

1.5. Workscope

The scope of this project is to re-evaluate the sustainable yield of 16 Hawaii basal aquifers for which deep monitoring well data are available. The re-evaluation of the sustainable yield of Pearl Harbor aquifer, which comprises three hydraulically connected

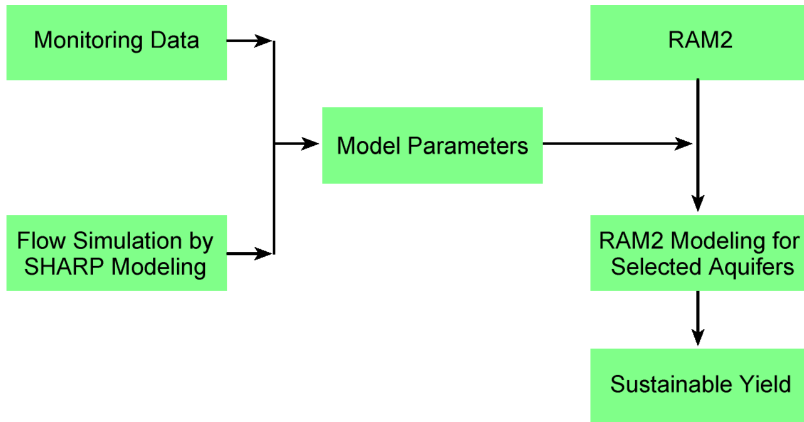


Figure 3. Research approach flowchart

aquifers on Oahu, is presented in a project report published by the Water Resources Research Center (Liu, 2006). The respective sustainable yields of Honolulu aquifer in southern Oahu, Iao aquifer in western Maui, and Kualapuu aquifer in central Molokai—as re-evaluated by RAM2 modeling—are presented in this report. The locations of these aquifers are shown in Figure 4.

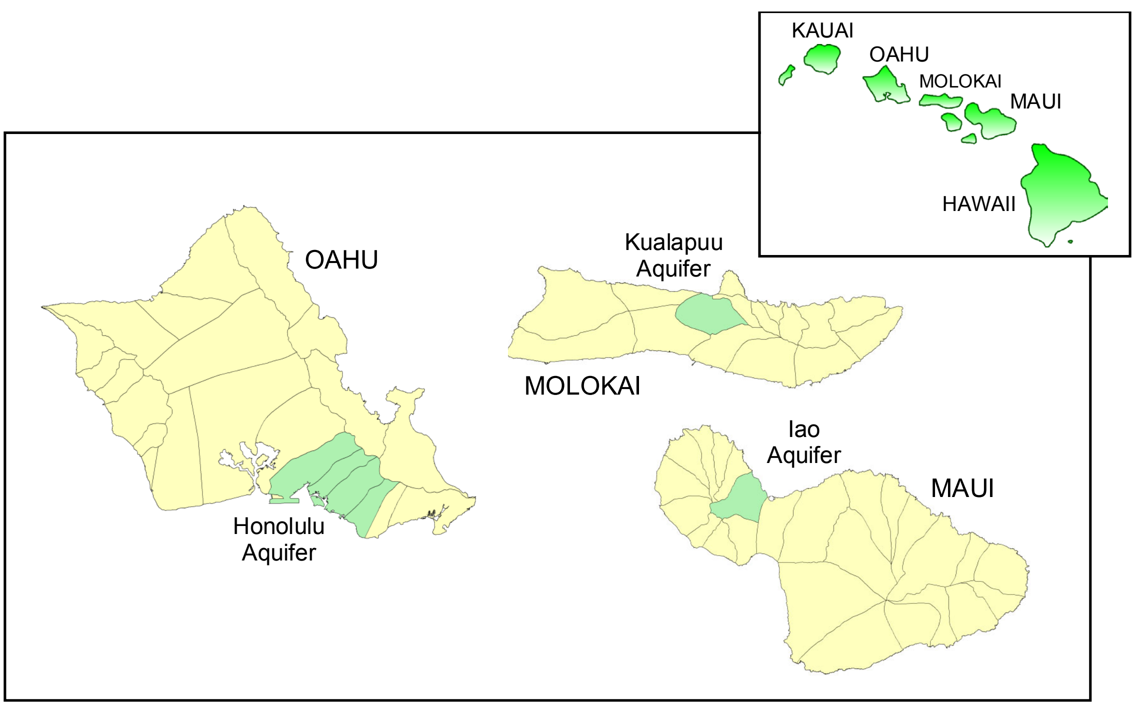


Figure 4. Locations of Hawaii basal aquifers investigated

In the case of the Honolulu aquifer, the respective sustainable yields of its four separate basal-water bodies (i.e., Moanalua, Kalihi, Beretania, and Kaimuki) are re-evaluated individually.

The sustainable yield of the remaining seven basal aquifers with deep monitoring well data cannot be re-evaluated by RAM2 at this time due to a lack of hydrogeologic data required for estimating more precisely the time-of-water travel from the upper boundary of the aquifer to each monitoring well.

2. METHODOLOGY

2.1. Mathematical Formulation of RAM2

RAM and RAM2 were both formulated conceptually by taking a basal aquifer as a completely stirred tank reactor (CSTR). RAM contains a sharp interface which separates the freshwater lens from the underlying saltwater (Figure 5a). In RAM2, however, a transition zone with varying salinity separates the upper and lower portions (Figure 5b).

As mentioned previously, RAM2 consists of a flow submodel and a salinity transport submodel. The flow submodel simulates the variation of the hydraulic head of a basal aquifer in response to pumping stress. The governing equation is

$$41SA\frac{dh}{dt} = \left[1 - \left(\frac{h}{h_0}\right)^2\right] I - D \tag{1}$$

where

S = storage coefficient

A = aquifer area

h = hydraulic head

I = rate of natural rainfall recharge

D = rate of net draft, or pumping minus irrigation return flow

As shown in Figure 5, the rate of leakage from a basal aquifer depends on the rate of natural recharge and the net draft, or $L = I - D$.

(a)

(b)

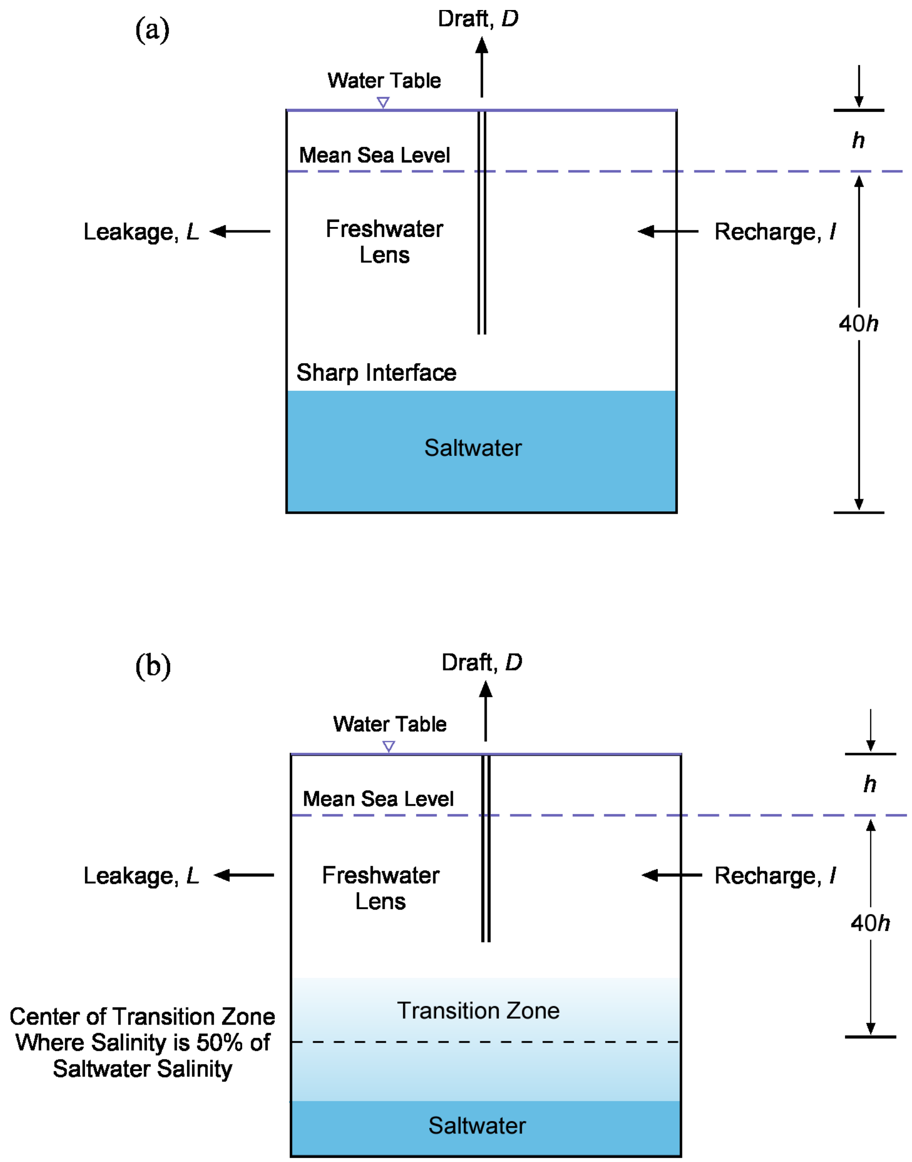


Figure 5. Conceptual formulation of (a) RAM as a CSTR with a sharp interface and (b) RAM2 as a CSTR with a transition zone

Equation (1) can be readily solved with an initial condition by giving the initial hydraulic head of the aquifer as h_0, or $t = 0$, $h = h_0$. The solution, which gives the variation of hydraulic head of a basal aquifer with respect to time, is shown as Equation (2) (Mink, 1981):

$$h_{i+1} = h_0 \sqrt{\frac{I-D}{I}} \left\{ \frac{\left[\sqrt{I-D} + \frac{h_i}{h_0}\sqrt{I}\right] \exp\left[\frac{2\sqrt{(I-D)I}\,(t_{i+1}-t_i)}{V_0}\right] - \sqrt{I-D} + \frac{h_i}{h_0}\sqrt{I}}{\left[\sqrt{I-D} + \frac{h_i}{h_0}\sqrt{I}\right] \exp\left[\frac{2\sqrt{(I-D)I}\,(t_{i+1}-t_i)}{V_0}\right] + \sqrt{I-D} - \frac{h_i}{h_0}\sqrt{I}} \right\} \tag{2}$$

The steady-state solution of Equation (2) is

$$\frac{h}{h_0} = \left(1 - \frac{D}{I}\right)^{1/2} \tag{3}$$

The head and draft relationships shown by Equation (3) are presented graphically in Figure 2.

The transport submodel simulates the salinity variation, in response to pumping stress, as a function of time in the transition zone of a basal aquifer. The governing equation of this submodel is in the form of the following steady-state advection-dispersion equation (Liu, 2006):

$$C(z) = \frac{C_0}{2}\left[1 - \operatorname{erf}\left(\frac{z}{\sqrt{4D_z\tau}}\right)\right] \tag{4}$$

where

C = salinity

C_0 = saltwater salinity

z = elevation above sharp interface

D_z = effective dispersion coefficient

τ = mean hydraulic residence time

A complete derivation of RAM2 is presented in a report by Liu (2006).

2.2. RAM2 Modeling

RAM2 is a viable management model for Hawaii basal aquifers. As shown in Figure 6, the sustainable yield of a basal aquifer can be determined by the integrated application of flow and transport submodels.

2.2.1. Calculating the Minimum Equilibrium Head

Acceptable source-water salinity in Hawaii is 1,000 μS/cm or less, which is about 2% of saltwater salinity. By setting the value of $C(z)/C_0$ in Equation (4) at 0.02, the corresponding value of z is ζ (Figure 6), or

$$0.02 = \frac{1}{2}\left[1 - \operatorname{erf}\left(\frac{\zeta}{\sqrt{4D_z\tau}}\right)\right] \tag{4a}$$

where

ζ = upper limit of a transition zone where water salinity is 2% of saltwater salinity

Forced draft from a basal aquifer causes freshwater head decline and transition zone expansion. Saltwater intrusion occurs when the upper limit of the transition zone in a basal aquifer reaches the bottom of a pumping well (Figure 6). The transport submodel can calculate the minimum equilibrium hydraulic head which must be maintained to prevent saltwater intrusion.

When water from a basal aquifer is pumped through a well, pumping stress causes a localized rising of the interface (Figure 7). This phenomenon, called upconing, is considered in the calculation of the minimum equilibrium hydraulic head by replacing the well depth with an effective well depth, which is defined as

$$\eta = h_{\text{well}} + h_{\text{upconing}} \tag{5}$$

where

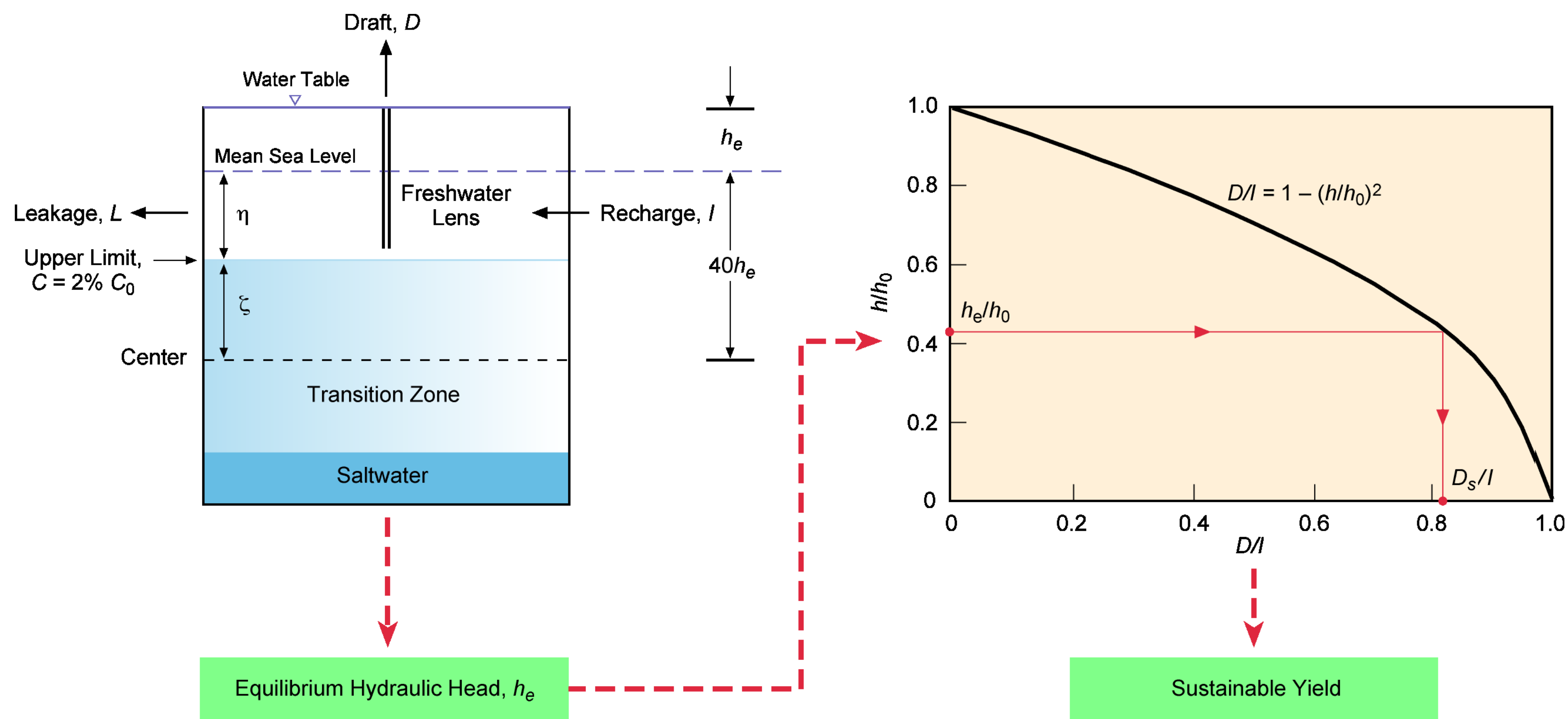


Figure 6. RAM2 modeling for determining the sustainable yield of a Hawaii basal aquifer

η = effective well depth used in the determination of the minimum equilibrium hydraulic head

h_{well} = well depth

h_{upconing} = upconing height

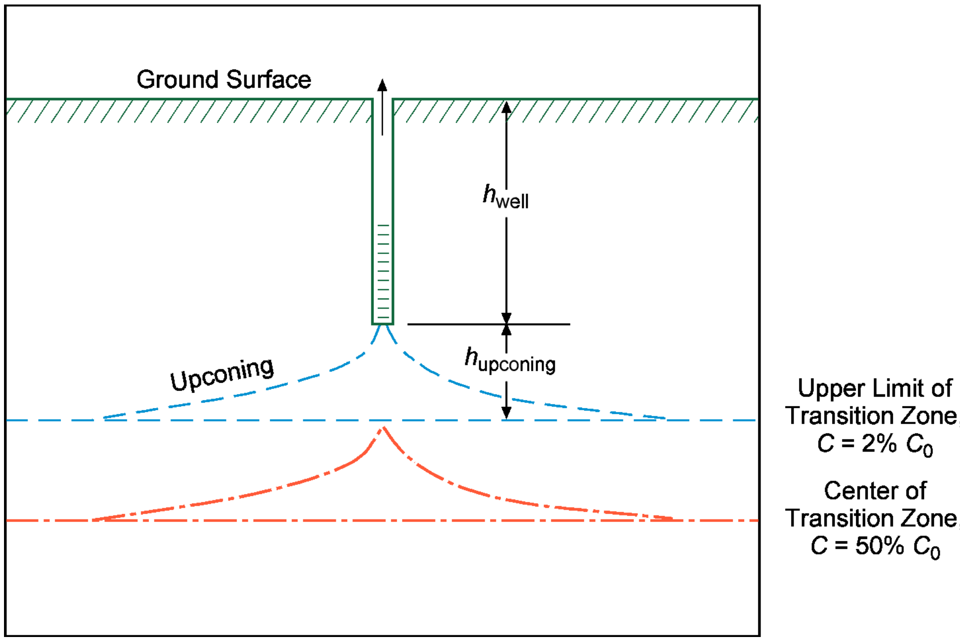


Figure 7. Upconing under a pumping well in a basal aquifer

If the center of a transition zone is taken as the location of the hypothetical sharp interface, the Ghyben–Herzberg relationship indicates that the minimum equilibrium hydraulic head (h_e) can be determined by the following equation (Figure 6):

$$h_e = \frac{\zeta + \eta}{40} \tag{6}$$

2.2.2. Calculating the Sustainable Yield

As shown on the right-hand side of Figure 6, RAM2 flow modeling gives a parabolic relationship of two dimensionless quantities of hydraulic head (h/h_0) and forced draft (D/I). It also indicates that the sustainable yield of a basal aquifer is the forced draft relative to the minimum equilibrium hydraulic head. The left-hand side of Figure 6 indicates that the minimum equilibrium hydraulic head can be calculated by RAM2 transport modeling. Thus, by combining the results of flow and transport modeling, the sustainable yield of a basal aquifer can be determined.

2.3. Model Parameters

The success of RAM2 modeling depends on how accurately model parameters are determined. Two principal model parameters of RAM2 are effective dispersion coefficient and mean hydraulic residence time. Multiplication of mean residence time by the effective dispersion coefficient, as shown in Equation (4), determines the elevation of particular salinity or the magnitude of salinity transport within the transition zone of a basal aquifer. An aquifer with a large dispersion coefficient or mean hydraulic residence time would contain a large transition zone and thus would be more susceptible to saltwater intrusion.

The dispersion coefficient of a heterogeneous aquifer is subject to macrodispersion and scale-dependent effects and thus cannot be satisfactorily estimated using data based on small-scale laboratory experiments (Gelhar et al., 1979; Neuman and Di Federico, 2003). Macrodispersion and scale-dependent effects have been investigated based on data collected by field tracer experiments. They have also been investigated by numerical experiments (Liu et al., 1991b; Azimi-Zonooz and Liu, 1994). These research efforts indicated that values of the dispersion coefficient of a particular aquifer have to be determined by field-scale tracer experiments (Gelhar et al., 1992). In these experiments, a conservative tracer was released from an injection well. A conservative tracer is not subject to any chemical or biological reaction and thus any change in its concentration is due to hydrodynamic dispersion only. Tracer breakthrough curves based on observations at monitoring wells located downstream from the injection well can therefore be analyzed to estimate dispersion coefficient. However, no successful field tracer study has been reported for Hawaii basal aquifers due to complications in conducting these experiments.

In this study, a natural field tracer method for estimating effective dispersion coefficients of Hawaii basal aquifers was developed. It is essentially a method of field tracer analysis using salinity as a natural conservative tracer. With this method, the effective dispersion coefficient at each deep monitoring well is estimated by a curve-fitting process that compares the salinity profile observed with that calculated using Equation (4b):

$$C(z) = \frac{C_0}{2}\left[1 - \operatorname{erf}\left(\frac{z}{\sqrt{4D_z t_R}}\right)\right] \tag{4b}$$

Equation (4b) is similar to Equation (4), except that the mean residence time (τ) is replaced by the time-of-water travel (t_R). Time-of-water travel is defined as the time for a water particle to travel from the upper boundary of a basal aquifer to the well.

The success of the natural field tracer method depends on (1) how accurately the salinity profile is measured at a deep monitoring well and (2) how accurately the time-of-water travel to the monitoring well is determined. In this project, time-of-water travel to each deep monitoring well in Iao and Kualapuu aquifers is determined by modeling analysis using SHARP (Essaid, 1990). Details of the time-of-water travel determination is shown in the Appendix.

Mean hydraulic residence time is the time a typical water particle would stay in a basal aquifer and is defined as V/Q, where V is the effective freshwater volume in a basal aquifer and Q is the aquifer outflow. Effective freshwater volume can be calculated as $V = 41(h)(n_0)(A)$, where h is the average hydraulic head, n_0 is the aquifer effective porosity, and A is the aquifer area. Under steady-state conditions, the outflow from a basal aquifer equals its leakage (Figure 5a).

2.4. Research Procedure

RAM2 is used to evaluate the sustainable yield of basal aquifers using the following procedure (Figure 8):

1. Deep monitoring well data concerning hydraulic heads and salinity profiles are used to determine the values of model parameters, i.e., dispersion coefficient and mean hydraulic resident time.
2. Values of model parameters are introduced to the transport submodel, or Equation (4).
3. The minimum equilibrium hydraulic head is calculated.
4. The ratio of the minimum equilibrium hydraulic head and the initial hydraulic head is then used by the flow submodel, or Equation (3), to determine the sustainable yield.

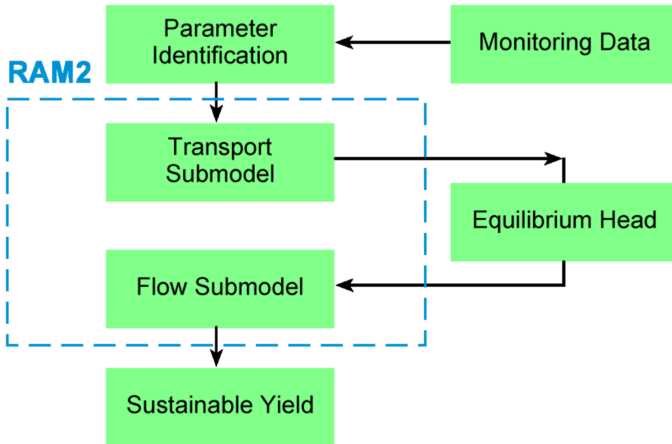


Figure 8. RAM2 modeling flowchart

3. SUSTAINABLE YIELD OF IAO AQUIFER ON MAUI

3.1. Aquifer Description

The Iao aquifer system (State Code 60102) is located on the west side of the western Maui mountain area, where it covers an area of about 24.7 mi^2 (Meyer and Presley, 2001). The area includes agricultural land for cultivation of sugarcane, pineapple, macadamia nuts, and other crops, as well as some urban land. The aquifer provides more than 60% of the domestic water supply for the island. In 1995 and 1996, water withdrawal from the aquifer approached the sustainable-yield value of 20 mgd. This value, which was established by the Hawaii Commission on Water Resource Management, is included in the Hawaii Water Resources Protection Plan (CWRM, 1990).

Over the last few years, water level in Iao aquifer has declined below that predicted by RAM (Oki and Meyer, 2001). At the same time, the transition zone between the freshwater and saltwater has risen, and the salinity of water pumped from wells in the aquifer has increased. As a result, forced draft from the aquifer has been cut back to a level below the established sustainable yield to prevent further decline of water levels and saltwater intrusion.

Table 3 lists basic data for existing production wells in Iao aquifer, including average pumping rate, top and bottom elevations, and initial hydraulic head. Adding the average pumping rate for all wells yields a total of 17.158 mgd for the aquifer system.

Table 3. Basic Data for Production Wells in the Iao Aquifer System

Well Name	Well No.	Year Drilled	Average Pumping Rate (mgd)	Elevation at Top of Well (ft)	Elevation at Bottom of Well (ft)	Initial Hydraulic Head (ft)
Shaft 33	5330-05	1946	5.025	32	-280	26.0
Mokuhau 1	5330-09	1953	1.396	353	-247	23.3
Mokuhau 2	5330-10	1953	0.532	353	-247	21.5
Mokuhau 3	5330-11	1967	2.599	354	-251	
Kepaniwai	5332-05	1973	0.840	713	413	677.0
Waiehu Hts. 1	5430-01	1975	0.195	337	-338	18.0
Waiehu Hts. 2	5430-02	1975	1.095	337	-206	18.0
Waihee 1	5431-02	1976	0.863	498	-182	13.6
Waihee 2	5431-03	1976	1.841	493	-150	
Waihee 3	5431-04	1981	2.772	493	-156	14.7
Total			17.158			

NOTE: Kepaniwai well is in a high-level aquifer.

3.2. Hydrogeology

The boundaries of Iao aquifer are the ridge south of Waihee River and north of Kalepa Gulch extending from the coast to the summit of West Maui Mountain, the crest of West Maui Mountain, the ridge north of Waikapu Stream extending from the crest to the isthmus, and the southern divide of Iao Stream to Kahului Bay (Figure 9).

Fresh groundwater in Iao aquifer occurs as dike-impounded groundwater and as a freshwater-lens system. The dike area receives high rainfall and provides much of the recharge to the freshwater-lens system, which is located downgradient. Also, perched freshwater occurs where saturated low-permeability sedimentary deposits or low-permeability volcanic rocks overlie unsaturated rocks. Perched water bodies are generally small and separated from the freshwater-lens system (Meyer and Presley, 2001).

Since most of the forced draft takes place in the basal aquifer, this study focuses only on the freshwater-lens section in the dike-free area.

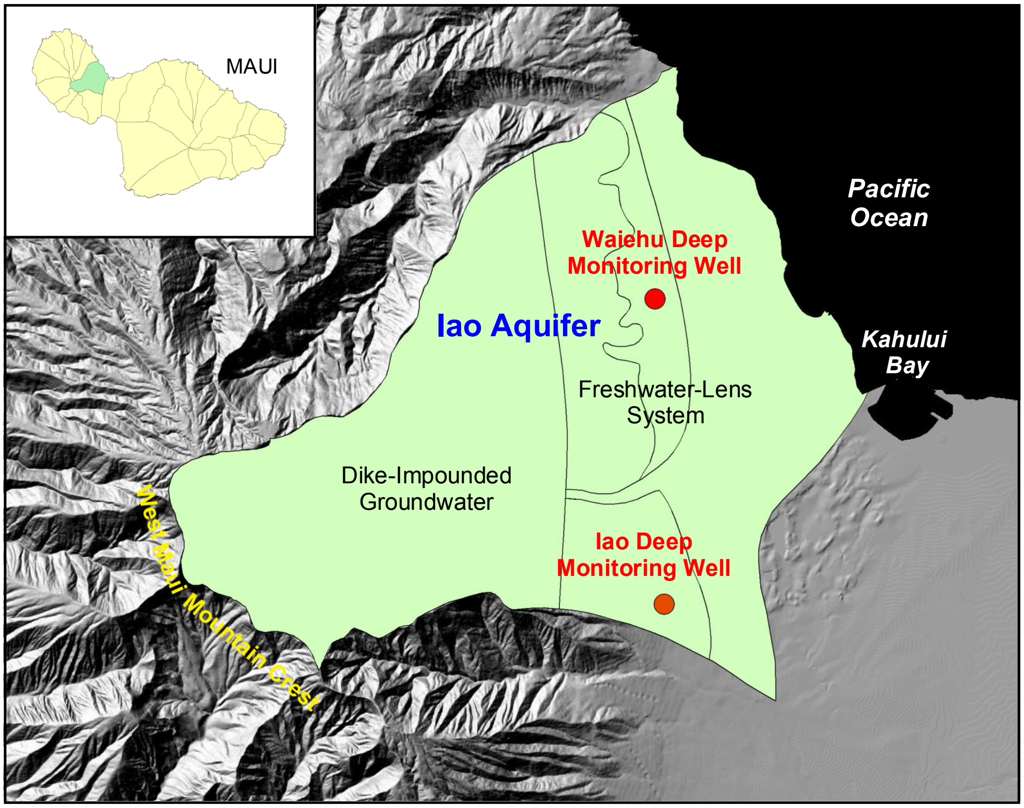


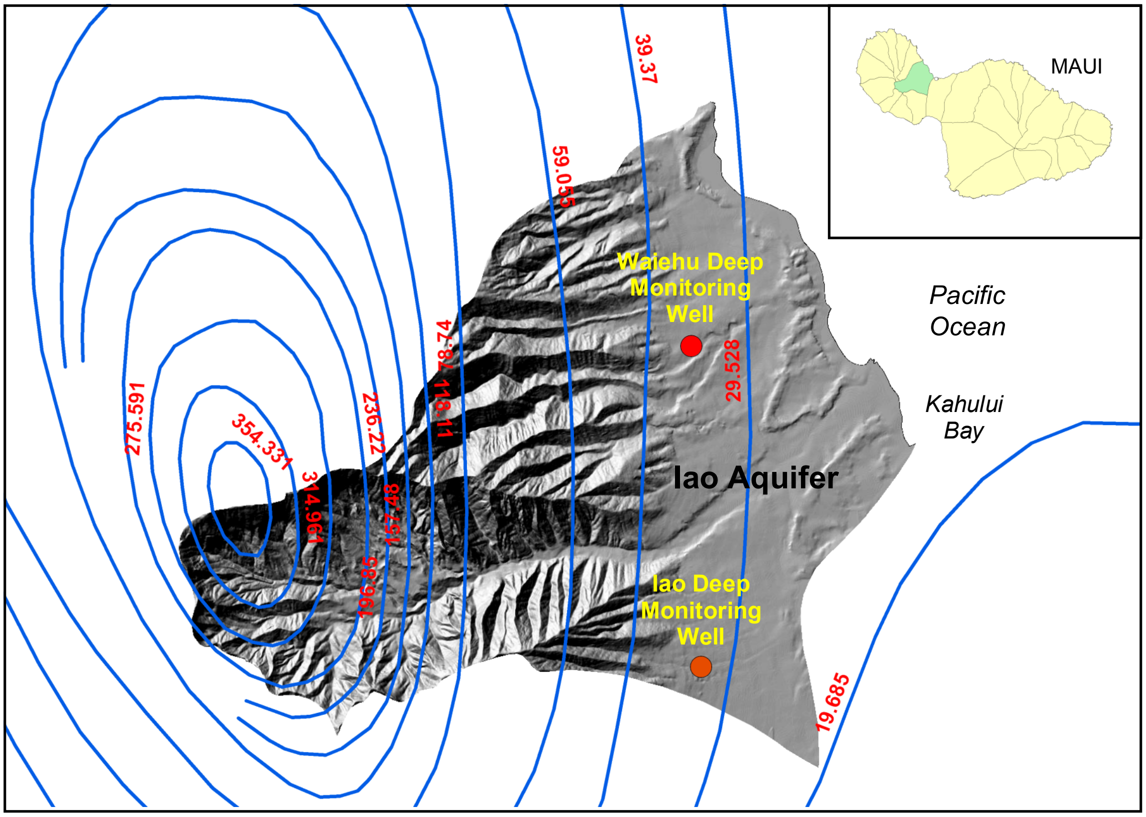
Figure 9. Boundaries and locations of deep monitoring wells in Iao aquifer, Maui, Hawaii

Groundwater flow studies in Hawaii basal aquifers are based on regional values of hydraulic conductivity and effective porosity. Groundwater flow near individual wells is controlled by local hydrogeologic characteristics, which are not well understood (Lau and Mink, 2006). The regional values of hydraulic conductivity in the Iao aquifer system were estimated at 1,000 ft/d for the dike-free area and 10 ft/d for the caprock area. The effective porosity of the dike-free area was estimated at 0.1.

3.3. Hydrologic Balance and Natural Recharge

Water enters the freshwater-lens system of Hawaii basal aquifers by direct infiltration of precipitation and irrigation water, seepage from irrigation ditches and streams where the water table is below the streambed, and inflow from upgradient dike-impounded water.

The recharge of Iao aquifer is mainly dependent on rainfall, although a small amount is contributed by agricultural return irrigation flow and fog drip. As shown in Figure 10, rainfall distribution varies, with the highest rainfall occurring in the mountain area and the lowest in



NOTE: Rainfall in inches.

Figure 10. Mean annual rainfall distribution at Iao aquifer, Maui, Hawaii

the coastal area. Mean annual rainfall at Puu Kukui is about 355 in. This number declines rapidly moving toward the ocean, such that the shoreline mean annual rainfall is 30 in. or less.

If the Iao aquifer is in a hydrologic balance, the rate of recharge would equal the rate of forced draft plus the rate of leakage to the ocean. The recently recorded average rate of forced draft from the aquifer is about 19 mgd. The rate of leakage from the aquifer to the ocean is estimated at 11 mgd. Note that the return irrigation flow from the agricultural area and from ditches, which is taken as part of the recharge, was estimated at 1.5 mgd by Meyer and Presley (2001). Therefore, the rate of natural recharge is about 28 mgd (CWRM, 1990).

The natural recharge of Iao aquifer for the period from 1986 to 1995 was estimated at 35 mgd by Shade (1997b). More recently, the natural recharge was estimated as high as 44 mgd (Engott, 2006). Such discrepancies are mainly due to different assumptions concerning various components in the hydrologic balance equation. These assumptions should be further investigated. In this project, natural recharge is estimated conservatively at 28 mgd.

3.4. Transport Parameters

3.4.1. Mean Hydraulic Residence Time

The mean hydraulic residence time of water particles in Iao aquifer, simulated as a CSTR, is defined as $\tau = \frac{V}{Q}$. The most recent estimation of leakage from Iao aquifer to the ocean is 11 mgd, or 1.470×10^6 ft^3/d. The average hydraulic head is 10.95 ft. The present volume of freshwater can be calculated as $V = 41(h)(n_0)(A) = 41(10.95)(0.1)(4.96 \times 10^8) = 2.23 \times 10^{10}$ ft^3. Therefore, mean hydraulic residence time of the Iao aquifer under current conditions can be estimated as

$$\tau = \frac{V}{Q} = \frac{2.23 \times 10^{10}}{1.470 \times 10^6} = 1,5170\,\text{d}$$

3.4.2. Dispersion Coefficient

SHARP was applied to simulate the hydraulic head contour and flow net of Iao aquifer under the current rate of forced draft. The simulated contour and flow net were then used to calculate time-of-water travel from the upper boundary of the aquifer to the deep monitoring wells.

In the Iao aquifer, the upper boundary is defined as the inferred boundary between the dike-impounded groundwater and the dike-free freshwater lens (Figure 9). Figure 11 shows the hydraulic head contour simulated by SHARP. The simulated flow net, consisting of head contour lines and flow lines which are perpendicular to the contour lines, is illustrated in Figure 12. From this simulated flow net, the time-of-water travel from the aquifer's upper boundary to Waiehu deep monitoring well was estimated at 2,913 days and to Iao deep monitoring well at 5,334 days.

Once the time-of-water travel is determined, the dispersion coefficient can be estimated by comparing calculated and observed salinity profiles. Figures 13 and 14 show calculated and observed salinity profiles that give the minimum mean square error for both Waiehu and

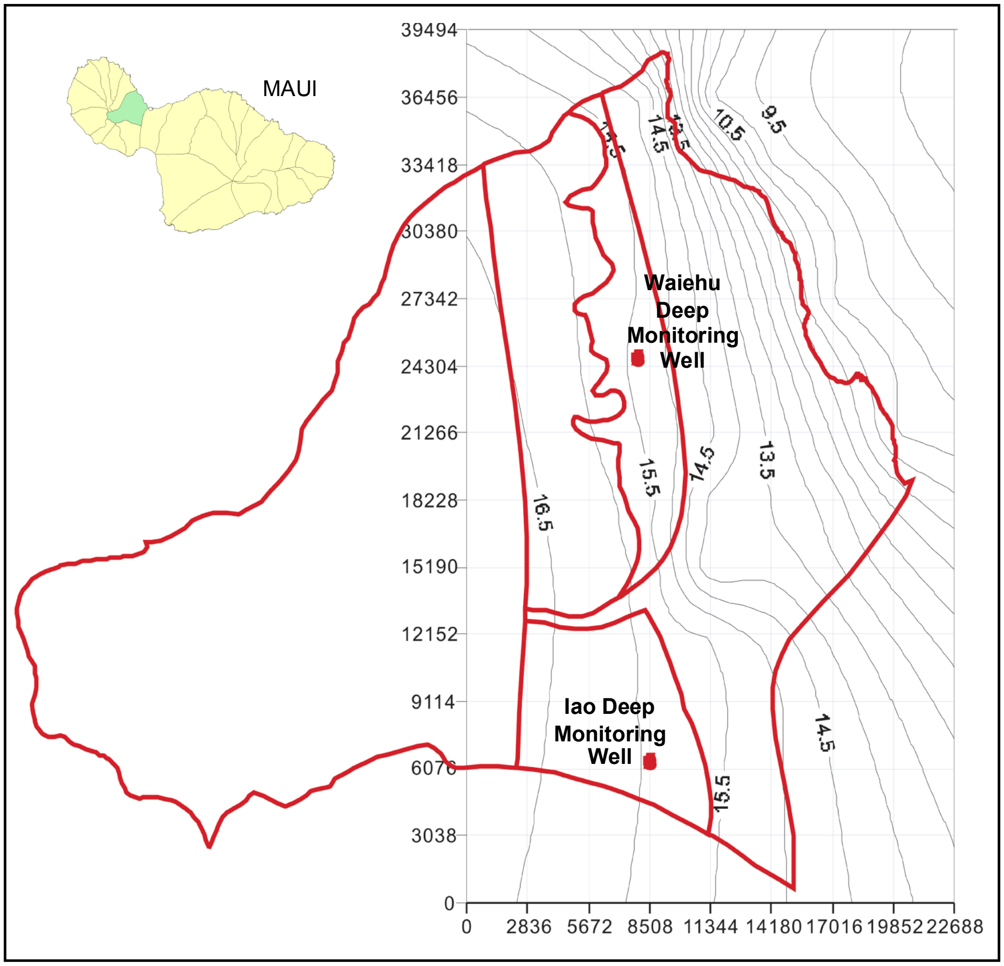


Figure 11. Hydraulic head contour of Iao aquifer, simulated by SHARP

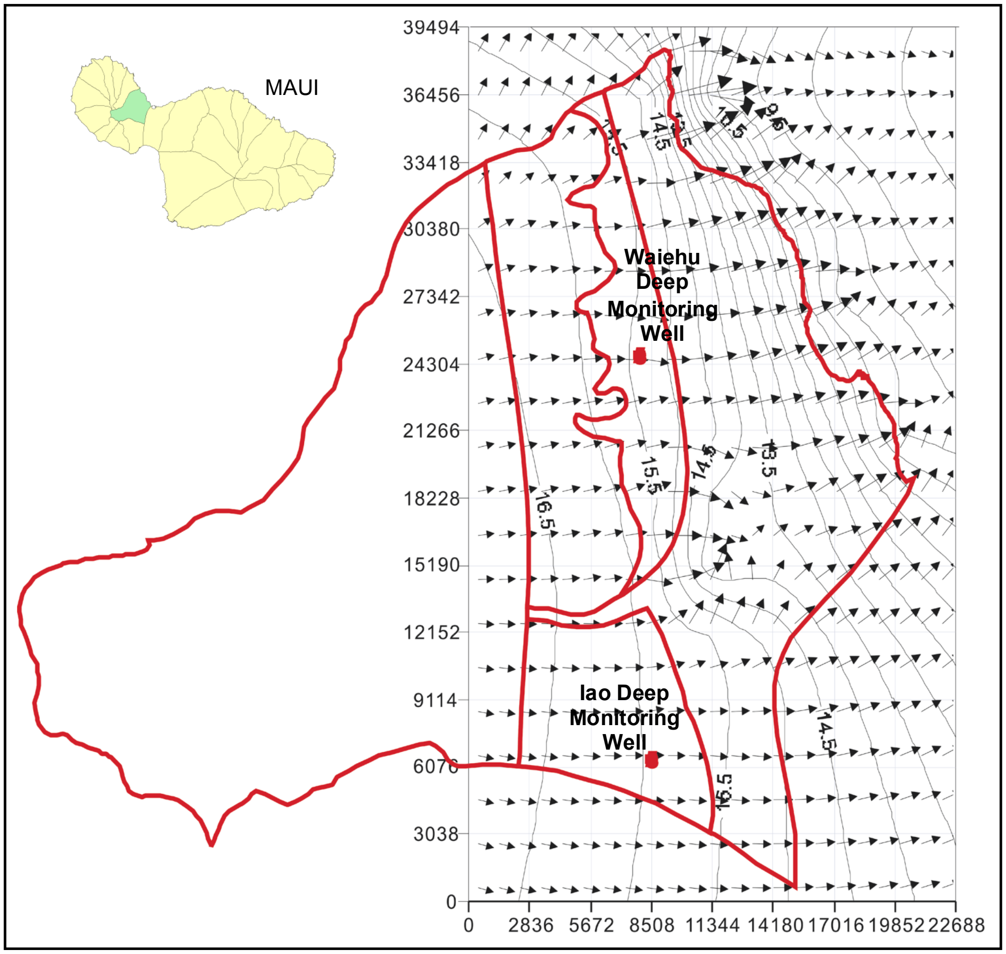


Figure 12. Groundwater flow net of Iao aquifer, simulated by SHARP

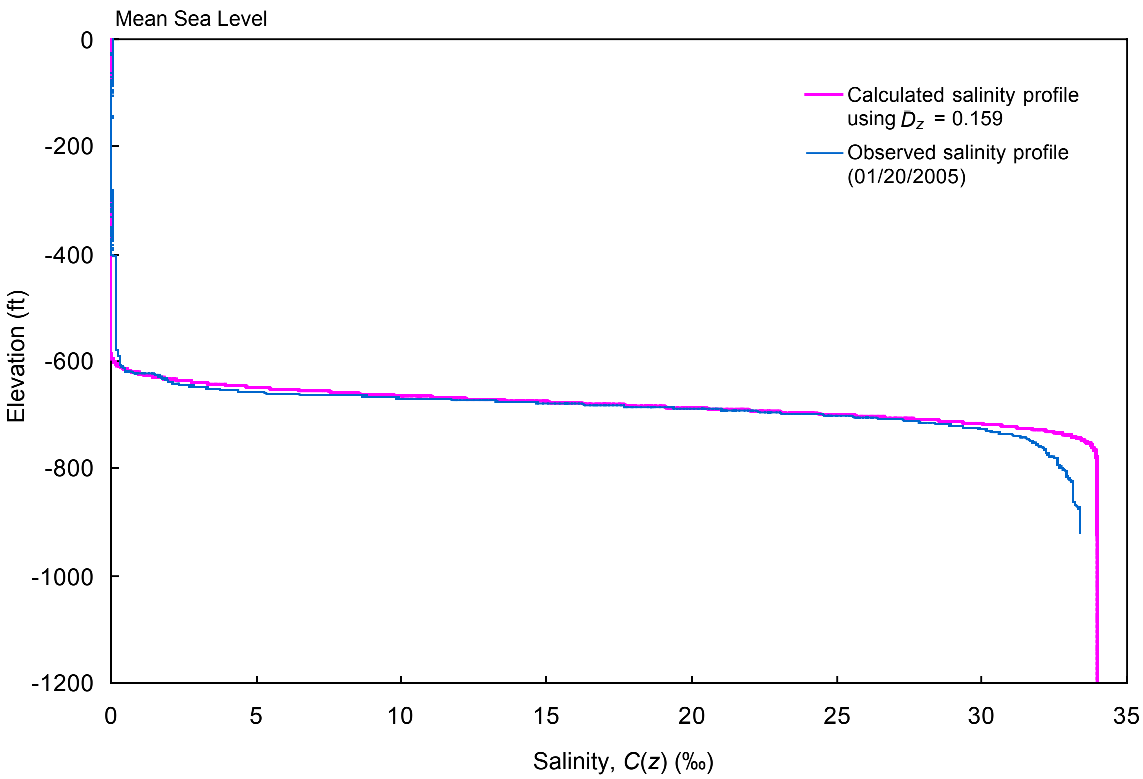


Figure 13. Estimation of dispersion coefficient by observed data fitting for Waiehu deep monitoring well

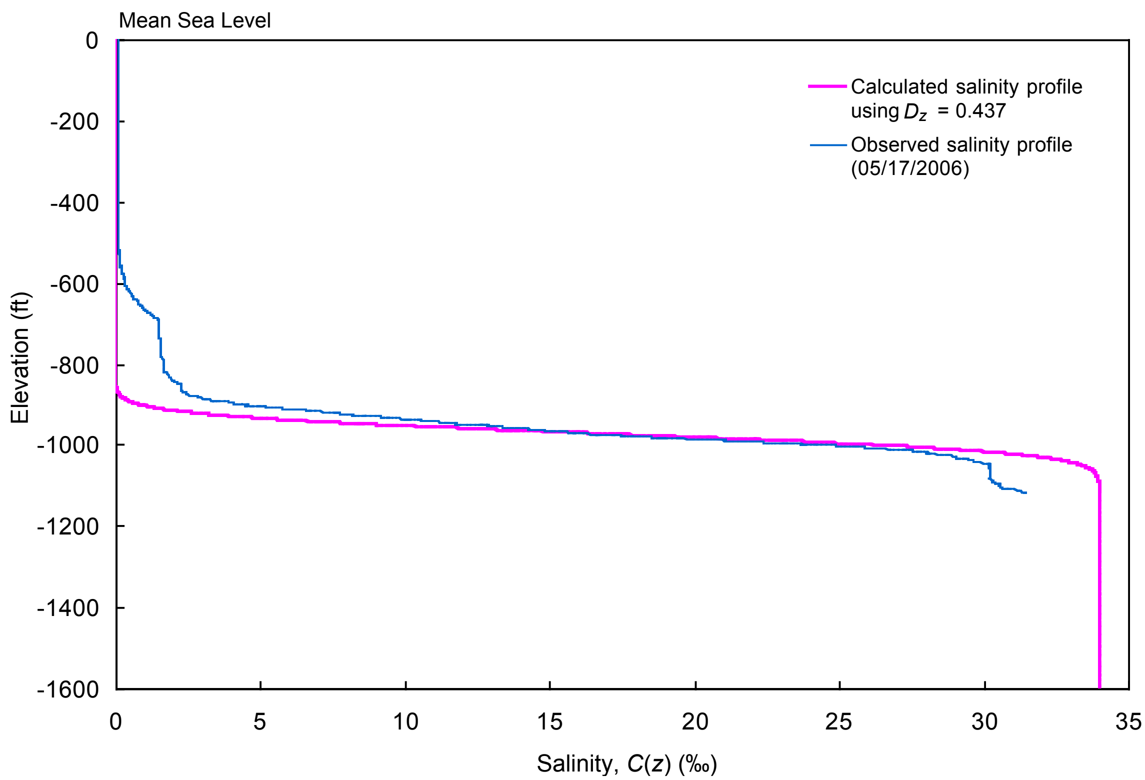


Figure 14. Estimation of dispersion coefficient by observed data fitting for Iao deep monitoring well

Iao deep monitoring wells. As indicated by Equation (4b), the shape of a calculated salinity profile changes with the value of the dispersion coefficient. The estimated dispersion coefficient results in a profile which has the minimum mean square error with the observed profile.

The dispersion coefficients estimated by using data taken at Waiehu deep monitoring well and Iao deep monitoring well are 0.159 ft^2/d and 0.437 ft^2/d, respectively. The average of these two values, or 0.298 ft^2/d, is taken as the effective dispersion coefficient of Iao aquifer.

3.5. Sustainable Yield Estimation

With values of dispersion coefficient and mean hydraulic residence time, the upper limit of the transition zone in Iao aquifer can be calculated by Equation (4a). In this case, D_z = 0.298 ft^2/d and τ = 15,170 d, so ζ = 196 ft.

Waiehu Height 1 pumping well, the deepest pumping well in Iao aquifer, penetrates 338 ft below mean sea level (Meyer and Presley, 2001). The upconing effect is about 100 ft (Liu, 2006). Thus, the effective well depth below mean sea level is 438 ft.

As shown in Figure 15, the minimum equilibrium head calculated by Equation (6) gives the following results: $h_e = (196 + 438)/40 = 15.85$ ft.

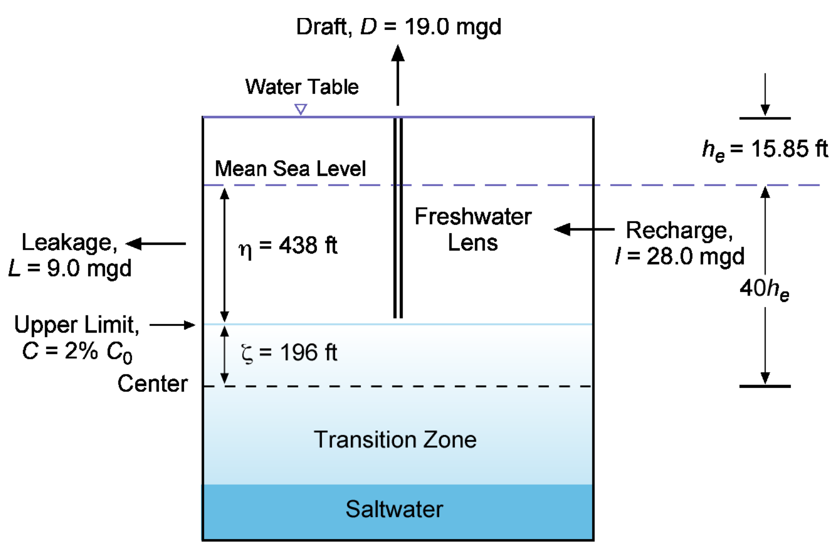


Figure 15. RAM2 transport submodeling for the determination of the equilibrium head of Iao aquifer

The initial average water level, or h_0, of Iao aquifer was estimated at 28 ft (CWRM, 1990). Therefore, the dimensionless head variable relative to the minimum equilibrium head, or h_e/h_0, is about 0.57. Based on the parabolic relationship between hydraulic head and forced draft derived by RAM, as illustrated in Figure 16, the dimensionless variable of forced draft is $D_s/I = 0.68$. The sustainable yield of Iao aquifer can then be determined by multiplying the dimensionless draft variable by the rate of natural recharge, or $D_s = 28\ (0.68) = 19.0$ mgd.

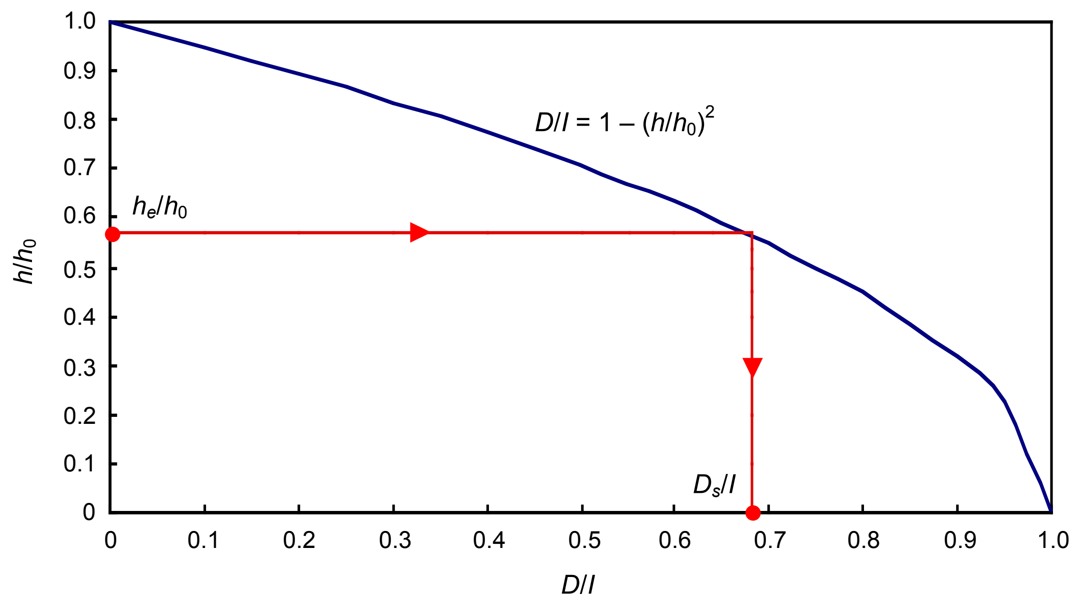


Figure 16. Determination of sustainable yield of Iao aquifer by RAM2

4. SUSTAINABLE YIELD OF KUALAPUU AQUIFER ON MOLOKAI

4.1. Aquifer Description

The island of Molokai was formed by shield volcanoes. Lavas of the East Molokai Volcano overlap the eastern flank of West Molokai Volcano, forming an elongate island with about 90 miles of coastline. The Kualapuu aquifer (State Code 40203) is located in the north-central part of the island (Figure 17). It covers an area of about 18.2 mi^2, including a dike-free freshwater lens of about 13.0 mi^2 (CWRM, 1990). The aquifer is the principal domestic water supply source for the island of Molokai.

Land use in the Kualapuu area is classified as urban, rural, and agricultural. Principal crops are sugarcane, pineapple, and macadamia nuts.

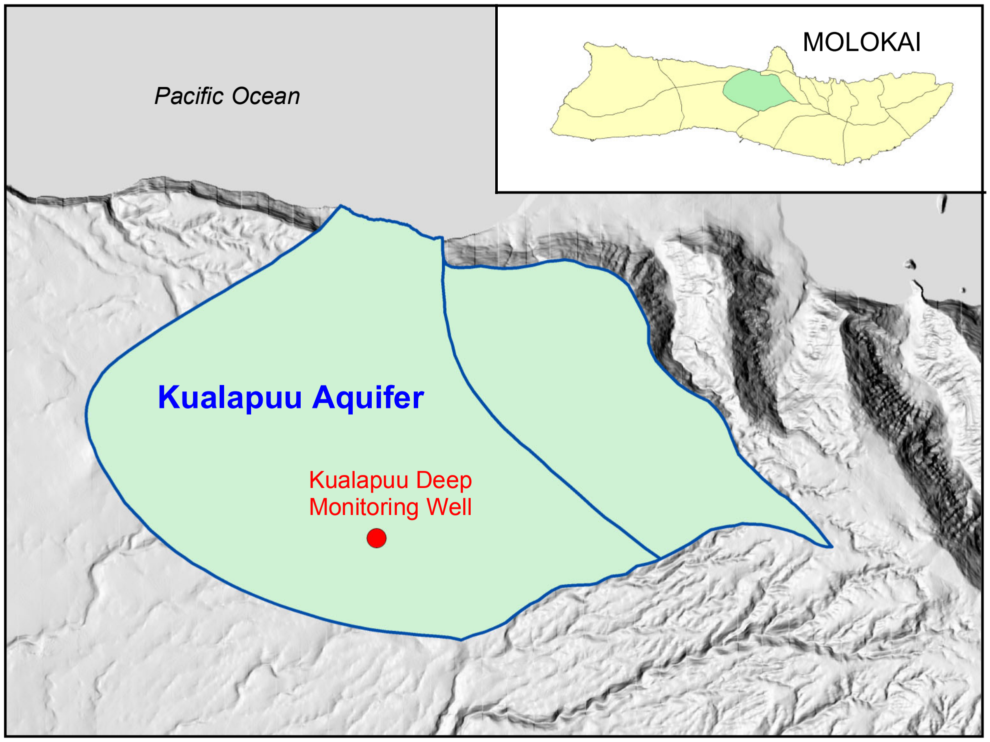


Figure 17. Boundaries and location of deep monitoring well in Kualapuu aquifer, Molokai, Hawaii

Five production wells (0801-01 to -03, 0901-01, and 0902-01) in the Kualapuu freshwater-lens system provide water for either irrigation or domestic use (Table 4). Wells 0801-01 and 0801-02 were drilled in 1948 and 1979, respectively, and Maui County well 0801-03 was drilled in 1987. Monthly mean withdrawal rates from these wells have remained below 1 mgd. Wells 0902-01 and 0901-01, drilled in 1946 and 1950, respectively, were originally used to irrigate pineapple fields in the Hoolehua Plain area. Well 0902-01 was abandoned in 1964, when water from the Molokai Irrigation System became available. Since 1976, water from well 0901-01 has been used for domestic and irrigation purposes in western Molokai. Prior to the completion of the Molokai Irrigation System tunnel, combined withdrawals from wells 0901-01 and 0902-01 varied seasonally from near zero to about 1 mgd. During 1996, the annual mean withdrawal rate from the four active wells in the Kualapuu area was 2.029 mgd (Oki, 1997).

As shown in Table 4, the maximum withdrawal rate is nearly 6 mgd (CWRM, 1990).

Table 4. Groundwater Withdrawal Rate of Wells in Kualapuu Aquifer

Well No.	Well Name	Withdrawal Rate (mgd) Maximum	Withdrawal Rate (mgd) Minimum	Initial Hydraulic Head (ft)	Elevation (ft)	Depth (ft)	Depth Below MSL (ft)	Purpose
0800-01	Kualapuu Deep	—	—	9.05	982	1,585	-600	Monitoring well
0801-01	Kauluwai 1	0.612	0	10.70	1,005	1,095	-90	Irrigation or domestic use
0801-02	Kauluwai 2	1.080	0	7.80	1,011	1,100	-89	Irrigation or domestic use
0801-03	Kaulapuu Mauka	1.440	0	11.70	1,037	1,136		
0901-01	Well 17	2.570	0	10.60	981	1,064	-83	Irrigation or domestic use
0902-01	Kualapuu	0	0	10.50	889	967	-78	Irrigation use (abandoned in 1964)
1000-01	Waipunahona Tunnel	0.290	0	—	1,550	—	—	—
Total		5.992						

NOTE: MSL = mean sea level.

4.2. Hydrogeology

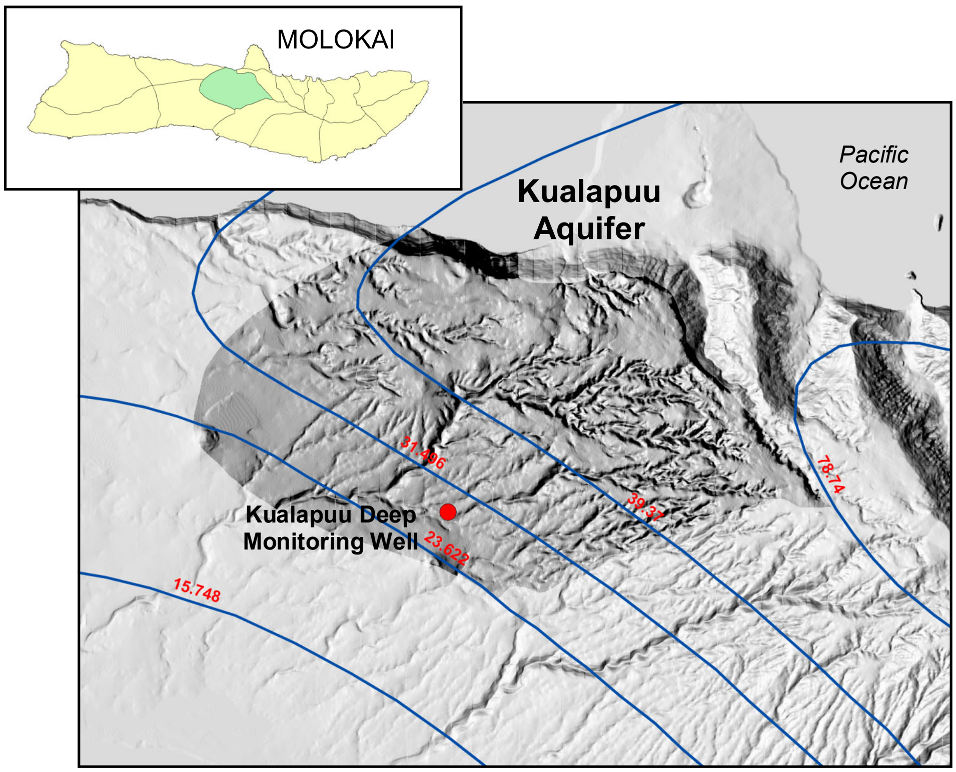
The island of Molokai is another volcanic doublet. The lavas of the East Molokai Volcano overlap the eastern flank of the older West Molokai Volcano to form a wide isthmus. Each volcano has two rift zones that are marked by cinder and spatter cones and by parallel dikes (Stearns and Macdonald, 1947). Emanating from a central caldera complex, the primary rift zones of the East Molokai Volcano trend northwest and east.

The Kualapuu aquifer is a freshwater-lens system located on the west side of East Molokai Volcano. The aquifer boundaries are shown in Figure 17. The freshwater in the aquifer includes dike-impounded groundwater, water atop saltwater in the freshwater-lens system, and perched water. This study focuses only on the freshwater-lens section in the dike-free area; the dike area and the perched water subsystems are ignored.

The hydraulic conductivity for the dike-free area was set at 1,000 ft/d and that for the low-permeability caprock area at 10 ft/d. The effective porosity of the dike-free area was estimated at 0.1.

4.3. Hydrologic Balance and Natural Recharge

On Molokai, maximum rainfall occurs at high altitudes and in areas with steep spatial gradients. Highest mean annual rainfall occurs in northeastern Molokai. Maximum mean annual rainfall of over 150 in. was recorded near the summit of East Molokai Volcano. Over West Molokai Volcano, mean annual rainfall is about 25 in., whereas along the coastal areas of southern and western Molokai, it is less than 16 in. Annual rainfall in the Kualapuu aquifer area varied from about 13 to 59 in. during the 1900 to 1993 period, as shown in Figure 18. The recharge of Kualapuu aquifer is mainly dependent on rainfall, with agricultural return irrigation and fog drip contributing a small amount.



NOTE: Rainfall in inches.

Figure 18. Annual rainfall distribution in Kualapuu aquifer area, Molokai, Hawaii

In the Hawaii Water Resources Protection Plan, the rate of natural recharge to the approximately 18.2 mi^2 Kualapuu aquifer is estimated at 9 mgd and the initial hydraulic head at 10 ft (CWRM, 1990). The initial hydraulic head of 10 ft is probably an underestimation. In a recent USGS study, results of measurements by the resistivity method indicated that the hydraulic head in the Kualapuu aquifer is greater than 10 ft (Oki, 2005). After many years of heavy pumping (see Table 4), a significant head decline is to be expected, so 13 ft was chosen as the initial head by this study. Further study of this parameter is necessary, however.

The average draft of Kualapuu aquifer was just over 2 mgd in 1996. Since then, it has been increasing. In this study, the average draft is estimated at 3 mgd. A recharge rate between 9 and 11 mgd was estimated in a USGS report (Shade, 1997a). This study selected 9 mgd as a conservative estimation of the natural recharge rate. Further study is needed for a more accurate estimation, however.

The total outflow from the aquifer, or leakage, calculated by a hydrologic balance is about 6 mgd.

4.4. Transport Parameters

4.4.1. Mean Hydraulic Residence Time

The most recent estimation of leakage from the aquifer to the ocean is 6 mgd, or 0.802 × 10^6 ft^3/d. The average hydraulic head of the aquifer is calculated as 8.6 ft, based on observed water level at five locations. The dike-free aquifer area is 13 mi^2, or 3.63 × 10^8 ft^2. The present volume of freshwater can be calculated as $V = 41(h)(n_0)(A) = 41(8.6)(0.1)(3.63 \times 10^8) = 1.28 \times 10^{10}$ ft^3. Therefore, mean hydraulic residence time of the Kualapuu aquifer under current conditions can be estimated as follows:

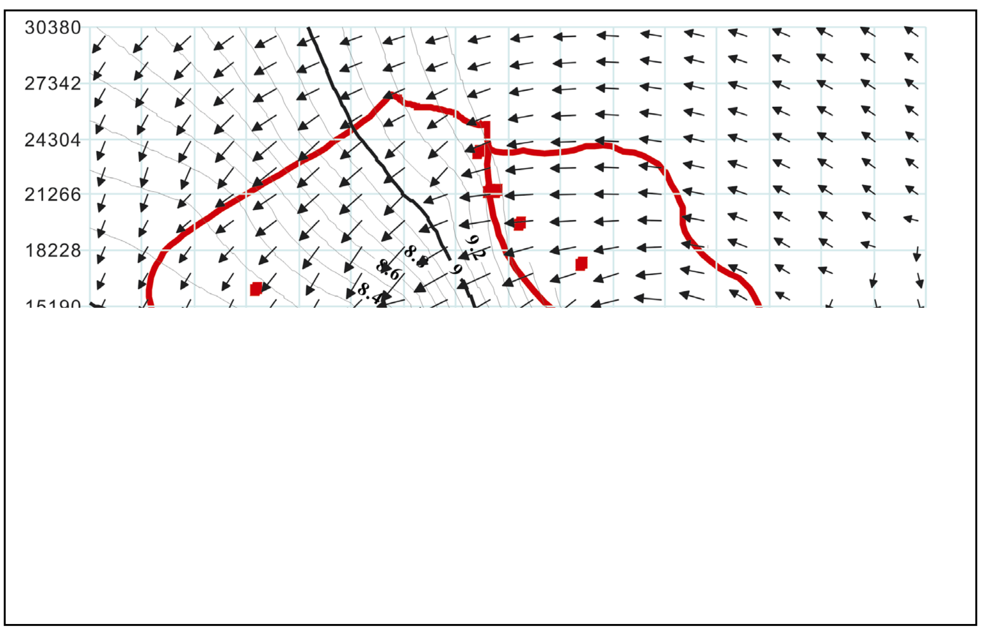
$$\tau = \frac{V}{Q} = \frac{1.28 \times 10^{10}}{0.802 \times 10^{6}} = 15,930\,\mathrm{d}$$

4.4.2. Dispersion Coefficient

SHARP was applied to simulate the freshwater flow in Kualapuu aquifer and to calculate time-of-water travel to monitoring wells. The simulated hydraulic head contour and flow net are shown in Figures 19 and 20, respectively. From the simulated flow net, the time-of-water travel from the aquifer's upper boundary to Kualapuu deep monitoring well was estimated at 1,894 days.

NOTE: Contours in ft.

Figure 19. Hydraulic head contour of Kualapuu aquifer, simulated by SHARP



The dispersion coefficient of Kualapuu aquifer can be estimated by comparing the salinity profile calculated using Equation (4b) with the observed profile (Figure 21). The estimate of 0.116 ft^2/d is taken as the effective dispersion coefficient of Kualapuu aquifer.

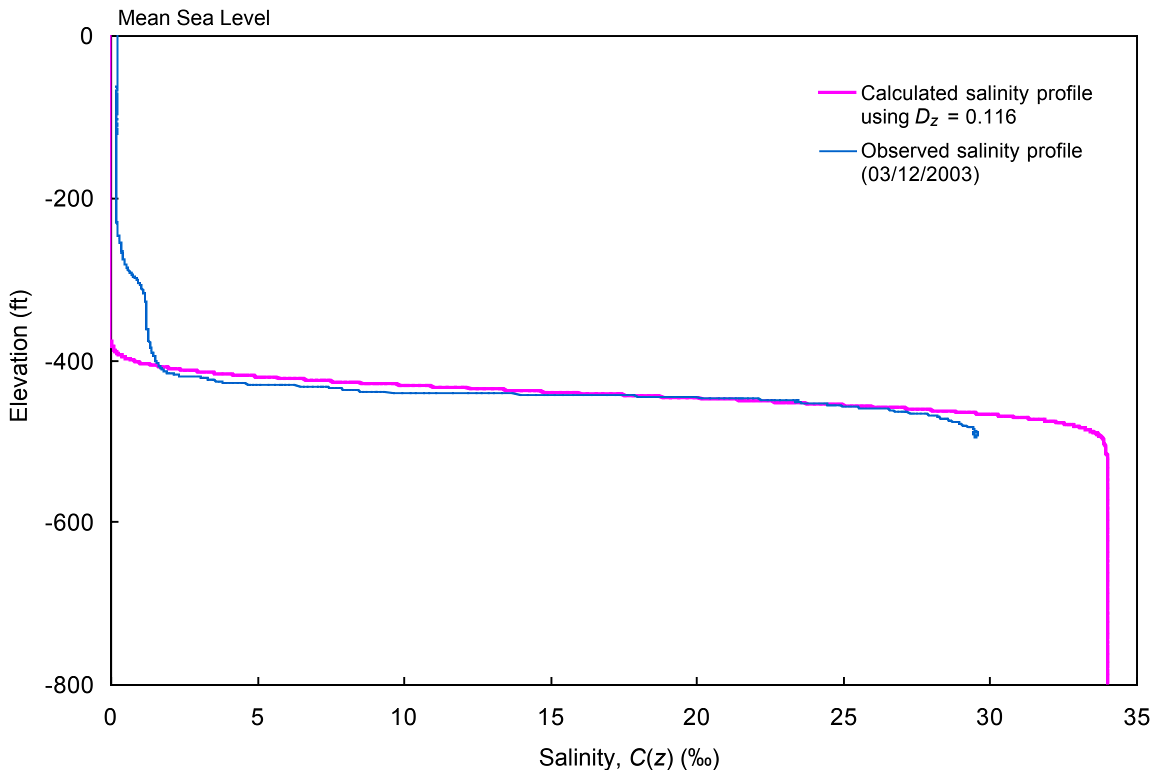


Figure 21. Estimation of dispersion coefficient by observed data fitting for Kualapuu deep monitoring well

4.5. Sustainable Yield Estimation

With values of dispersion coefficient and mean hydraulic residence time, the upper limit of the transition zone in Kualapuu aquifer can then be calculated by Equation (4a). In this case, $D_z = 0.116$ ft^2/d and $\tau = 15{,}930$ d, so $\zeta = 125$ ft.

The average depth of pumping wells in Kualapuu aquifer is about 120 ft, and the estimated upconing effect is 100 ft (Liu, 2006). Thus, the effective well depth below mean sea level is 220 ft.

As shown in Figure 22, the minimum equilibrium head calculated by Equation (6) gives the following result: $h_e = (125 + 220)/40 = 8.62$ ft.

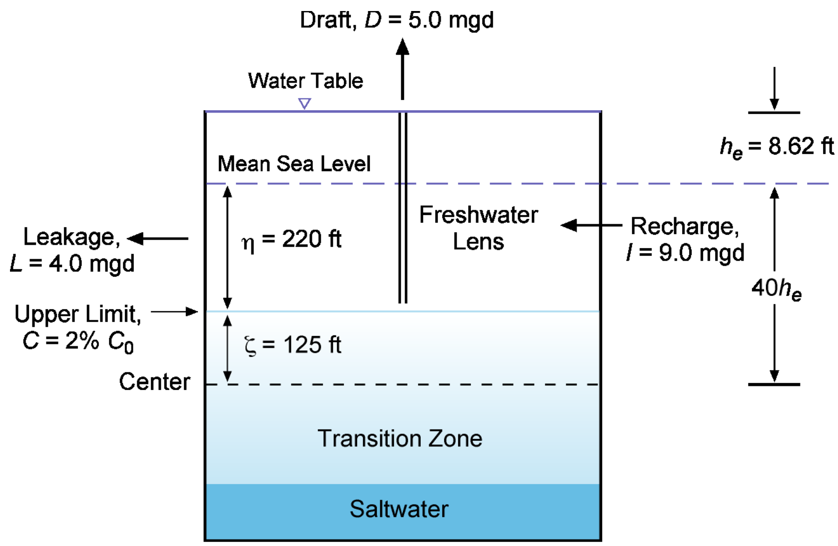


Figure 22. RAM2 transport submodeling for the determination of the equilibrium head of Kualapuu aquifer

The initial average water level, or h_0, of Kualapuu aquifer was estimated at 13 ft. Therefore, the dimensionless head variable relative to the minimum equilibrium head, or h_e/h_0, is about 0.66. Based on the parabolic relationship between hydraulic head and forced draft, the dimensionless variable of draft, or D_s/I, is about 0.56 (Figure 23). The sustainable yield of the aquifer can be determined by multiplying the dimensionless draft variable by the rate of natural recharge, or $D_s = 9\ (0.56) = 5.0$ mgd (Figure 23).

5. SUSTAINABLE YIELD OF BASAL-WATER BODIES IN HONOLULU AQUIFER ON OAHU

5.1. Aquifer Description

The Southern Oahu aquifer, located between the Waianae and Koolau mountain ranges, is the most productive groundwater body in Hawaii. For groundwater management purpose, this aquifer is treated as two separate basal aquifers—Pearl Harbor aquifer and Honolulu aquifer—even though they are hydraulically connected (Mink, 1980). Re-evaluation of the sustainable yield of Pearl Harbor aquifer by RAM2 modeling is detailed in a project report recently published by University of Hawaii Water Resources Research Center (Liu, 2006). This study presents the re-evaluation of the sustainable yield of Honolulu aquifer.

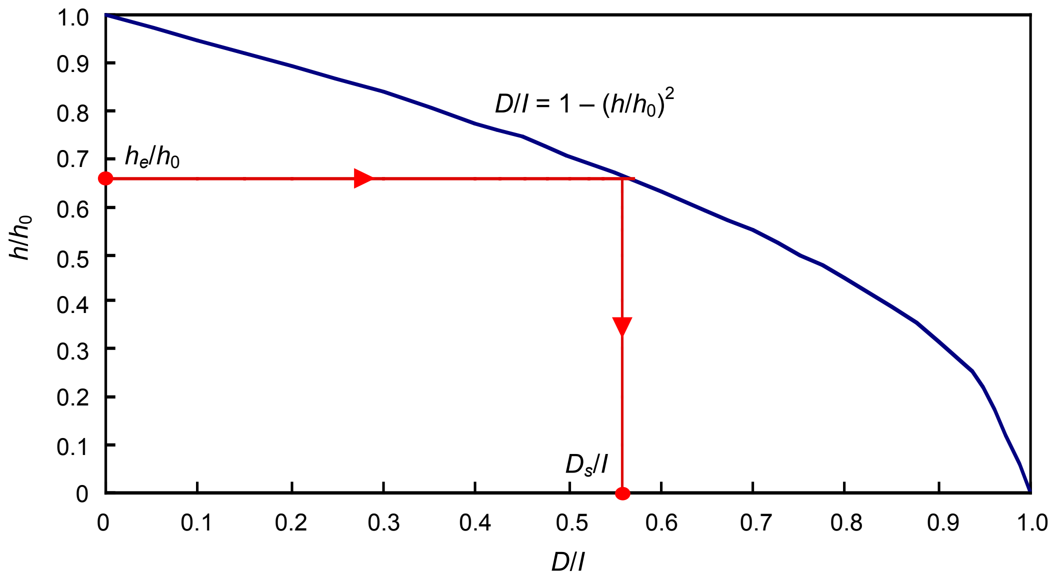


Figure 23. Determination of sustainable yield of Kualapuu aquifer by RAM2

The Honolulu aquifer has a total area of 61.6 mi^2 and is bounded on the east by southern Palolo Valley, on the north by the Koolau mountain range, on the west by Halawa Valley, and on the south by the Pacific Ocean (Visher and Mink, 1964) (Figure 24). This aquifer comprises four basal-water bodies or isopiestic areas, including Moanalua area (State Code 30104), Kalihi area (State Code 30103), Beretania area (State Code 30102), and Kaimuki area (State Code 30101) (Mink, 1980) (Figure 24). The boundaries of these areas have been modified by various past studies. More recent modifications, as shown in Figure 24, were suggested by USGS (Oki, 1998; Hunt, 1996). These modifications were adopted by Todd Engineers (2005) in its development of a groundwater management model of Honolulu aquifer. They were also adopted by this study.

Basic information of these four basal-water bodies is provided in Table 5. To the east of the Kaimuki area are two other basal groundwater bodies called Waialae and Wailupe–Hawaii Kai area I. These two groundwater bodies are located in a poorly defined zone of northeast-trending dikes (Takasaki and Mink, 1982). They are not included in this study.

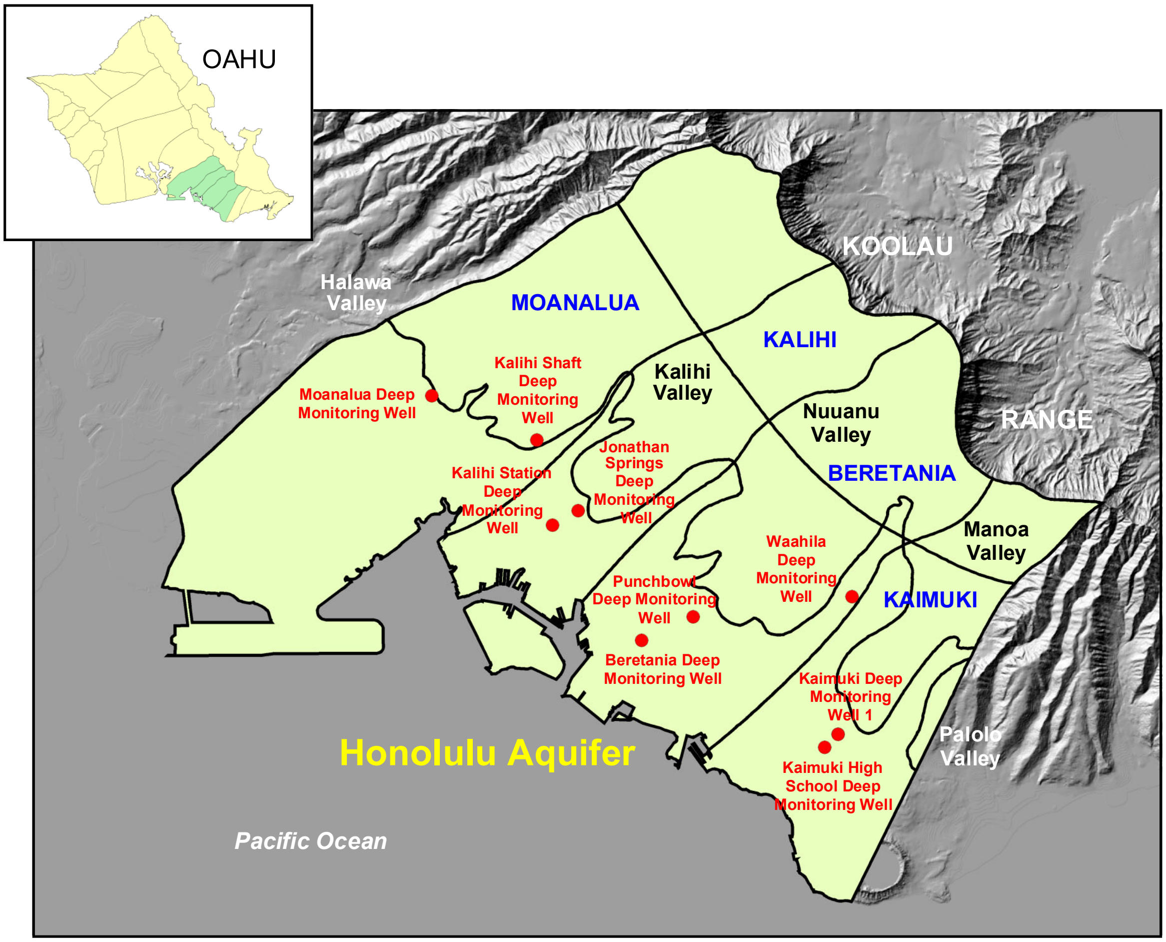


Figure 24. Boundaries of basal-water bodies and locations of deep monitoring wells in Honolulu aquifer, Oahu, Hawaii

Table 5. Basal-Water Bodies in the Honolulu Aquifer

Basal-Water Body	BWS Area Designation	State Code	Area (mi^2)	Area Boundaries
Moanalua	4	30104	10.92	North Halawa Valley to Kalihi Valley
Kalihi	3	30103	6.33	Kalihi Valley to Nuuanu Valley
Beretania	2	30102	8.62	Nuuanu Valley to Manoa Valley
Kaimuki	1	30101	4.43	Manoa Valley to Palolo Valley

NOTE: BWS = Board of Water Supply.

5.2. Hydrogeology

Lava flow formations composing the groundwater areas in the Honolulu aquifer are extremely permeable, with a hydraulic conductivity of about 1,500 ft/d. These groundwater areas are bounded on all sides by comparatively impermeable geologic formations. Dike-intruded lava flows form the boundary near the Koolau mountain range, whereas deep valley-fill alluvium in stream valleys forms lateral boundaries perpendicular to shore. Near the shoreline, these groundwater bodies are overlain by poorly permeable marine and terrestrial sediments, with a hydraulic conductivity of about 10 ft/d.

Deep valley-fill alluvium, which separates these groundwater areas, is rather impervious. With a hydraulic conductivity estimated by laboratory permeameter tests as between 0.083 and 0.128 ft/d (Wentworth, 1938), the alluvium essentially prevents exchange of freshwater between areas. However, a hydraulic connection exists between areas at depths below the alluvium; at these depths the aquifer is saturated with saltwater. These groundwater areas are sometimes referred to as “isopiestic” or “isopotential” areas because the hydraulic head in each area is significantly different from that in the adjacent area (Mink, 1980). Figure 25 is a cross-sectional view of the Moanalua groundwater area in Honolulu aquifer. It illustrates the lateral boundaries between one basal-water body and adjacent basal-water bodies.

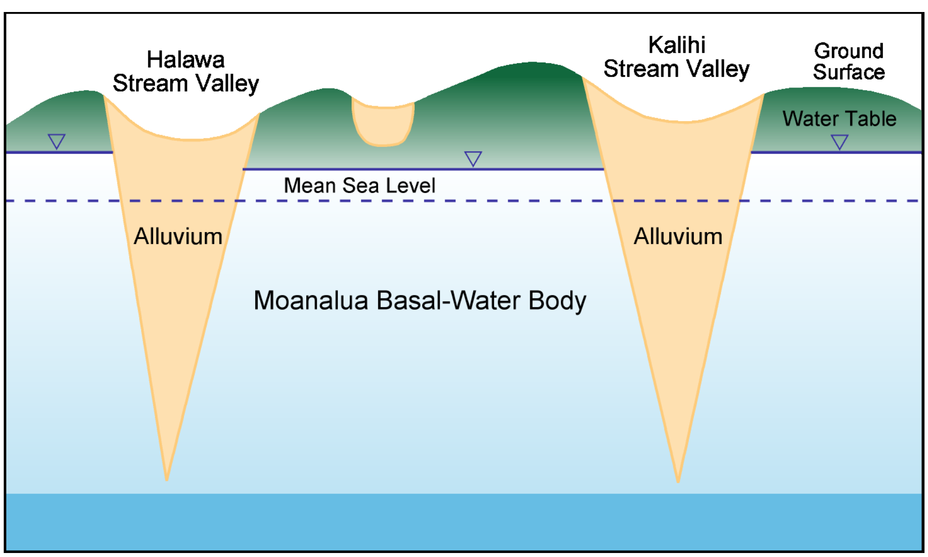


Figure 25. Geologic cross-section of Moanalua basal-water body in Honolulu aquifer

With little lateral movement across alluvial boundaries, groundwater in each basal-water body in Honolulu aquifer moves independently (Figure 26).

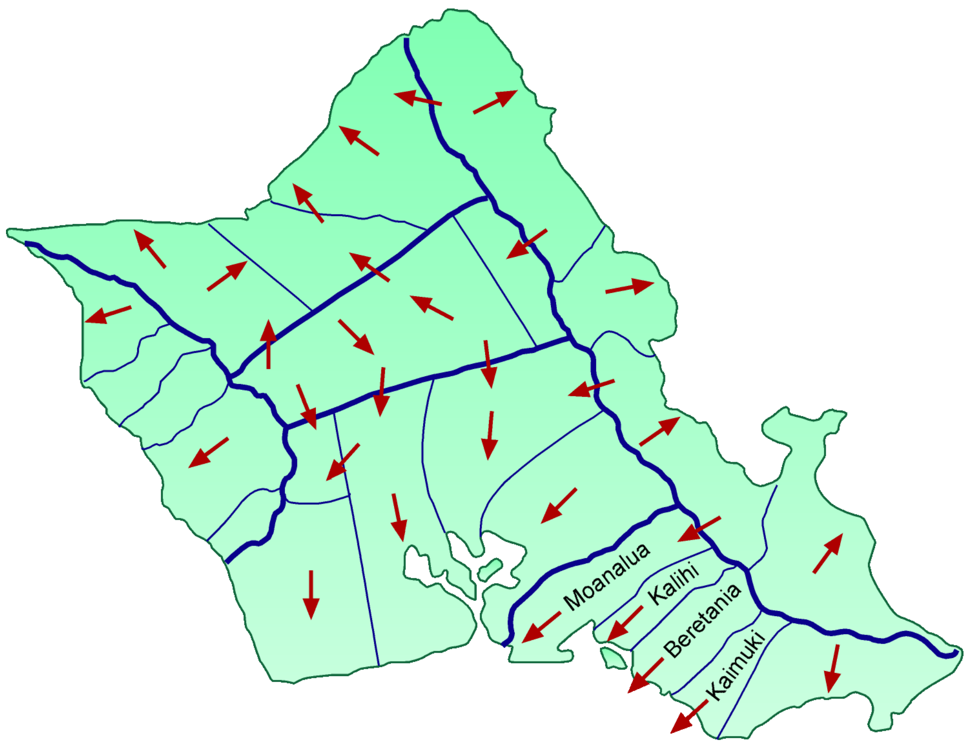


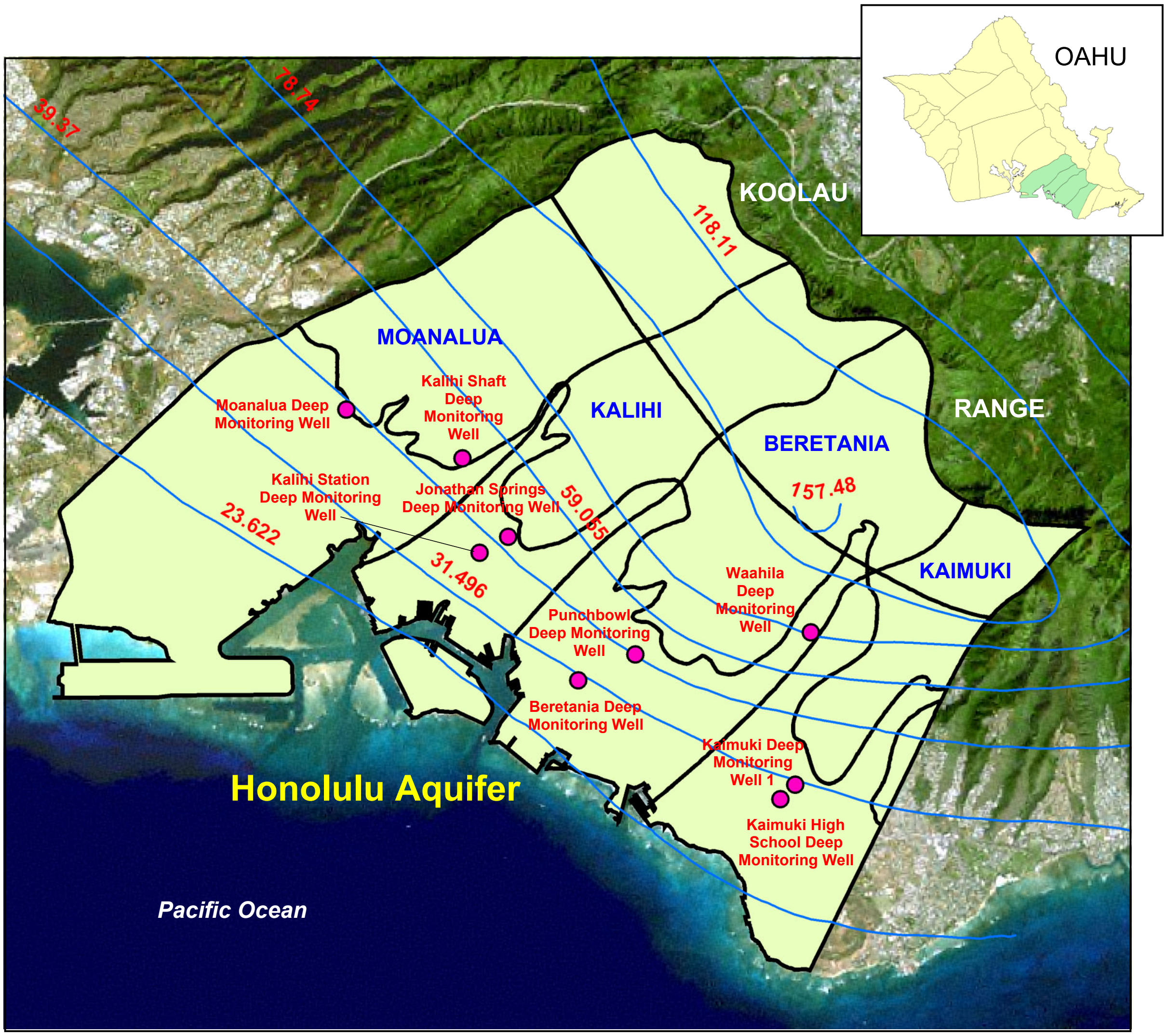
Figure 26. Water movement of Oahu aquifer systems (modified from Oki, 2005)

5.3. Hydrologic Balance and Natural Recharge

Spatial distribution of rainfall in the Honolulu aquifer area is shown by lines of equal rainfall isohyets in Figure 27. The rainfall varies with the highest amount (up to 158 in./yr) in the Koolau range area and the lowest in the coastal area.

Natural recharge is the difference between rainfall and the sum of average runoff and evapotranspiration. Table 6 gives the estimated rate of natural recharge provided in the Hawaii Water Resources Protection Plan (CWRM, 1990).

Todd Engineers (2005) conducted an analysis of natural recharge in the Honolulu aquifer based largely on data published in a technical report of the University of Hawaii Water Resources Research Center (Giambelluca, 1983). Generally, recharge rates estimated by Todd Engineers (2005) for the four basal-water bodies in Honolulu aquifer are higher than estimates provided in the Hawaii Water Resources Protection Plan (CWRM, 1990) (Table 6).



NOTE: Rainfall in inches.

Figure 27. Annual rainfall distribution of Honolulu aquifer, Oahu, Hawaii

Table 6. Estimated Rates of Natural Recharge for Basal-Water Bodies in Honolulu Aquifer

Basal-Water Body	Recharge[a] (mgd)	Recharge[b] (mgd)
Moanalua	24	31.1
Kalihi	12	11.6
Beretania	20	18.8
Kaimuki	6	7.2
Total	62	68.7

[a]Data from CWRM (1990).
[b]Data from Todd Engineers (2005).

Table 7 gives the hydraulic heads for each basal-water body in Honolulu aquifer, as determined prior to groundwater development and again in 2000. Forced draft in the aquifer has caused significant decline of hydraulic heads. The hydraulic head contour was simulated in Honolulu aquifer in 2000 by Todd Engineers (2005) by using a three-dimensional numerical model.

Table 7. Long-Term Water-Level Decline of Basal-Water Bodies in Honolulu Aquifer

Basal-Water Body	Hydraulic Head		Overall Decline
	Pre-Development[a]	Year 2000	
Moanalua	38	17.97	10.83
Kalihi	40	20.80	20.90
Beretania	42	20.96	22.54
Kaimuki	38	23.63	11.37

[a]Data from CWRM (1990).

5.4. Estimation of the Sustainable Yield of Moanalua Basal-Water Body

5.4.1. Aquifer Description

The Moanalua basal-water body (State Code 30104) is designated groundwater area 4 by the Honolulu Board of Water Supply. It is located between low-permeability valley-fill barriers underlying Halawa Stream and Kalihi Stream (Figure 24). It covers an area of 10.92 mi^2 (CWRM, 1990).

Rainfall distribution over the Moanalua basal-water body is illustrated by isohyetals in Figure 27. The average annual rainfall was estimated at 101 in. and the annual surface runoff and evaporation at 25 in. and 40 in., respectively. Thus, the rate of natural recharge is about 36 in./yr, or 24.0 mgd.

Data on active pumping wells in the Moanalua basal-water body and their respective depths below mean sea level are provided in Table 8. Total average draft for 1986 to 2000 was estimated at 16.4 mgd (Todd Engineers, 2005).

Pre-development average hydraulic head in the Moanalua basal-water body (h_0) was about 38 ft. Over the last 100 years, forced draft has caused the hydraulic head to decline to the current average head of about 18 ft (Table 7).

Table 8. Production Wells in Moanalua Basal-Water Body, 1986–2000

Well Name	Owner	Well No.	Number of Wells	GSE (ft)	Casing Depth (ft)	Well Depth (ft)	Elevation Below MSL (ft)
Moanalua	BWS	2153	10	36	150	300	-264
			11	35	150	300	-265
			12	35	185	335	-300
Fort Shafter	Private	2053	10	20	168	279	-259
			13	20	180	290	-270
			11	21	175	330	-309
Tripler	Private	2153	7	29	52	302	-273
			8	29	57	306	-277
Red Hill Shaft	Private	2254	1	200	N/A	210	-10
Hawaii Meat Company	Private	2053	9	22	449	607	-585
Honolulu Construction	Private	2053	5	20	347	471	-451
Sam Damon	Private	2153	2	20	79	289	-269
Kalihi Shaft	BWS	2052	8	160	129	154	6
Honolulu International Co.	Private	2154	1	14	103	294	-280
Average							-272

NOTE: GSE = ground surface elevation, MSL = mean sea level, BWS = Board of Water Supply, N/A = not applicable.

5.4.2. RAM2 Parameters

5.4.2.1. Mean Hydraulic Residence Time

The aquifer outflow is the difference between the rate of natural recharge and forced draft, or $Q = 24.0 - 16.4 = 7.6$ mgd, or 1.016×10^6 ft^3/d. The average hydraulic head is about 18.0 ft, and the area is 10.92 mi^2, or 3.04×10^8 ft^2. The present volume of freshwater can be calculated as $V = 41(h)(n_0)(A) = 41(18.0)(0.1)(3.04 \times 10^8) = 2.25 \times 10^{10}$ ft^3. Therefore, the mean hydraulic residence time of the water particles in Moanalua basal-water body under current conditions can be estimated as follows:

$$\tau = \frac{V}{Q} = \frac{2.25 \times 10^{10}}{1.016 \times 10^6} = 22{,}146 \text{ d}$$

5.4.2.2. Dispersion Coefficient

The effective dispersion coefficient of the Moanalua basal-water body is based on data taken at Moanalua deep monitoring well and Kalihi Shaft deep monitoring well (Figure 24). It is determined using a curve-fitting process that compares the observed and calculated salinity profiles. The estimated time-of-water travel of the water particle and an effective porosity of 0.1 are used in the calculation. The time required for water particles to travel from the upper boundary of the basal-water body to Moanalua deep monitoring well is 4,801 days and the time to Kalihi Shaft deep monitoring well is 2,760 days (Table 9).

The observed salinity profile for Moanalua deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 28. The estimated dispersion coefficient is 1.435 ft^2/d. The observed salinity profile for Kalihi Shaft deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 29. The estimated dispersion coefficient is 1.935 ft^2/d. However, it is noted that the lower half of the observed salinity curve for Kalihi Shaft deep monitoring well shows that salinity remained at about 25,000 mg/l at 850 ft below mean sea level, indicating possible measurement errors. Thus, instead of using the average of the two deep monitoring wells combined, the dispersion coefficient of Moanalua deep monitoring well, or 1.435 ft^2/d, is used as the effective dispersion coefficient for the Moanalua basal-water body.

Table 9. Water Particle Time-of-Water Travel to Deep Monitoring Wells in Moanalua Basal-Water Body

Parameter	Monitoring Well	
	Moanalua Deep	Kalihi Shaft Deep
High water head (ft)	22.5	22.8
Low water head (ft)	15.6	15.9
Hydraulic conductivity, K (ft/d)	1,000	1,000
Travel distance, L (ft)	18,200	13,800
Hydraulic gradient (ft)	0.00038	0.0005
Velocity, u (ft/d)	0.379	0.500
Porosity, P	0.10	0.10
True velocity, u' (ft/d)	3.79	5.00
Time-of-water travel, t_R (d)	4,801	2,760

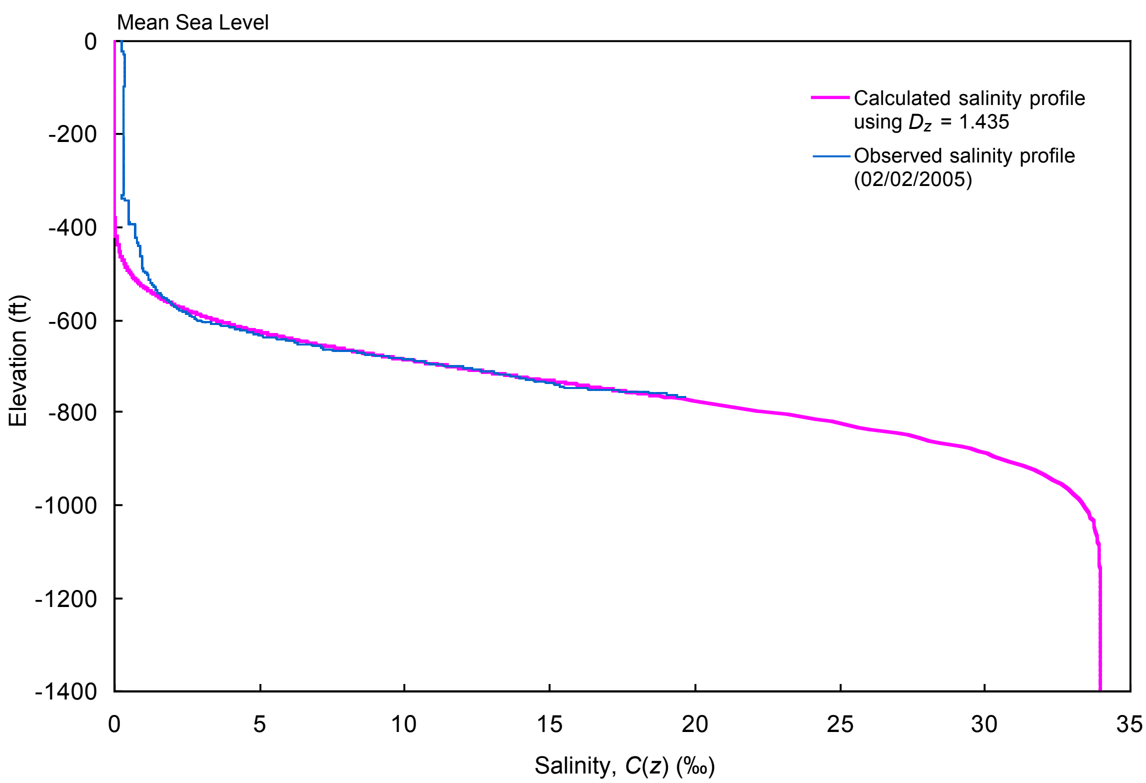


Figure 28. Estimation of dispersion coefficient by observed data fitting for Moanalua deep monitoring well

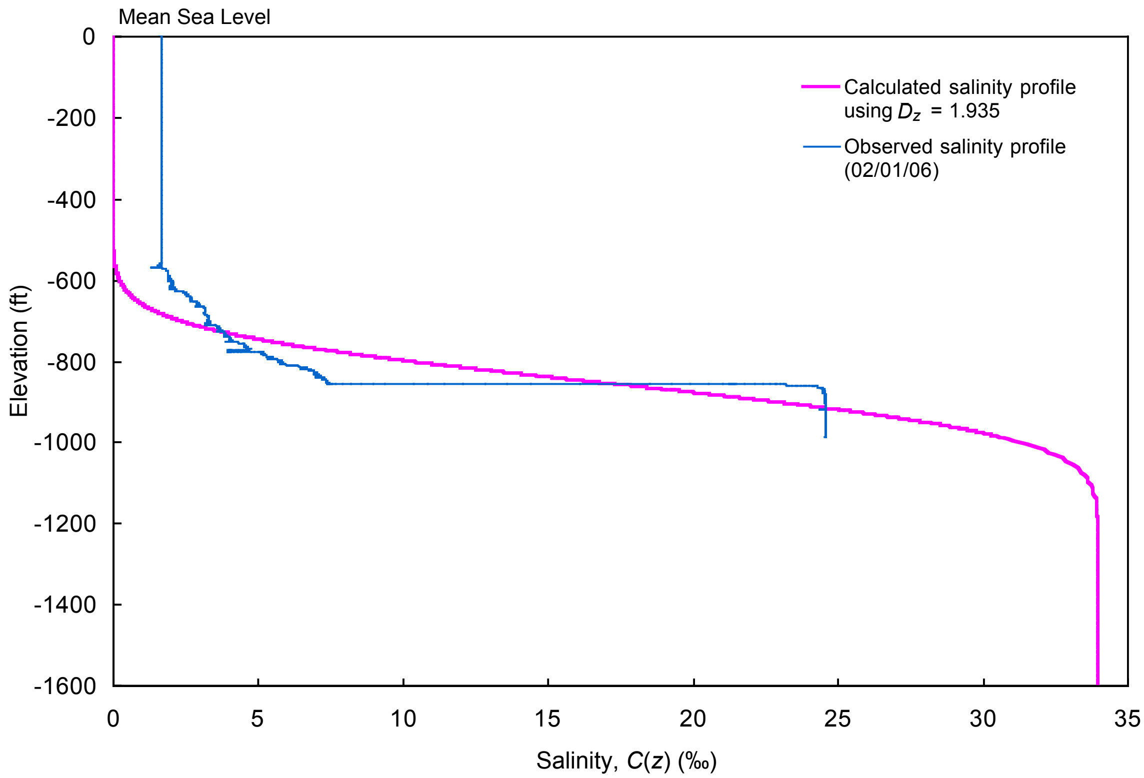


Figure 29. Estimation of dispersion coefficient by observed data fitting for Kalihi Shaft deep monitoring well

5.4.3. Sustainable Yield Determination

The sustainable yield of the Moanalua basal-water body can be determined using RAM2. The minimum equilibrium hydraulic head is calculated using the RAM2 transport submodel. The result is divided by the initial average water level to get the dimensionless variable of the minimum equilibrium hydraulic head.

The minimum equilibrium hydraulic head calculated by the RAM2 transport submodel for Moanalua basal-water body is illustrated in Figure 30. Values for effective dispersion coefficient (D_z) and mean hydraulic residence time (τ) are used to calculate the upper limit of the transition zone (ζ) using Equation (4a). In this case, $D_z = 1.435$ ft^2/d and $\tau = 22{,}150$ d, so $\zeta = 517$ ft. To determine the effective well depth below mean sea level, the well depth and the upconing height are added (Equation (5)). The average depth of production wells in the Moanalua basal-water body is about 272 ft below mean sea level, and the upconing height is

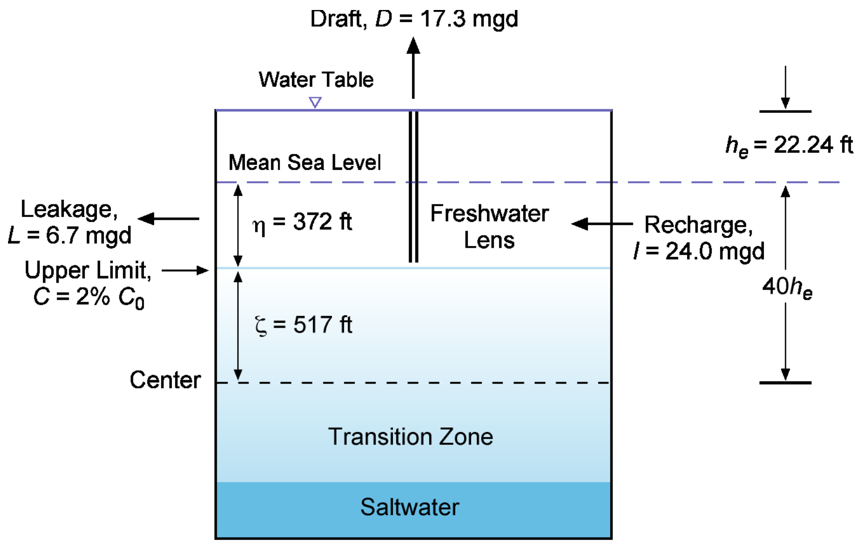


Figure 30. RAM2 transport submodeling for the determination of the equilibrium head of Moanalua basal-water body

about 100 ft (Liu, 2006). Thus, η = 372 ft. Values for both the upper limit of the transition zone and the effective well depth below mean sea level are used to calculate the minimum equilibrium head (Equation (6)). The result is as follows: $h_e = (517 + 372)/40 = 22.24$ ft.

Dividing the minimum equilibrium hydraulic head by the initial average water level, estimated at 42 ft (CWRM, 1990), gives the dimensionless variable of the minimum equilibirum hydraulic head. Thus, $h_e/h_0 = 22.24/42 = 0.53$. Based on the parabolic relationship between hydraulic head and forced draft derived by RAM, as illustrated in Figure 31, the dimensionless variable of forced draft is determined as $D_s/I = 0.72$.

The final step in determining the sustainable yield of a basal-water body using RAM2 is to multiply the rate of natural recharge by the dimensionless variable of draft. Thus, for Moanalua basal-water body the sustainable yield is calculated as $D_s = 24 \times 0.72 = 17.3$ mgd.

5.5. Estimation of the Sustainable Yield of Kalihi Basal-Water Body

5.5.1. Aquifer Description

The Kalihi basal-water body (State Code 30104) is designated groundwater area 3 by the Honolulu Board of Water Supply. It is located between low-permeability valley-fill

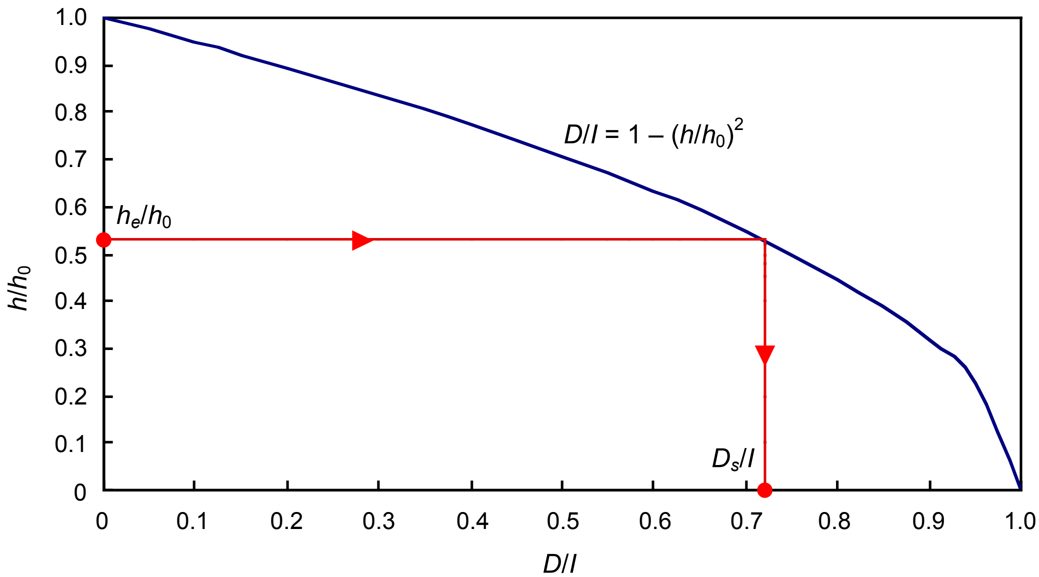


Figure 31. Determination of sustainable yield of Moanalua basal-water body by RAM2

barriers underlying Kalihi Stream and Nuuanu Stream (Figure 24). It covers an area of 6.33 mi^2.

Rainfall distribution over the Kalihi basal-water body is illustrated by isohyetals in Figure 27. The average annual rainfall was estimated at 90 in. and the annual surface runoff and evaporation at 20 in. and 40 in., respectively. Thus the rate of natural recharge is about 30 in./yr, or 12.0 mgd.

Data on active pumping wells in the Kalihi basal-water body and their respective depths below mean sea level are provided in Table 10. Total average draft for 1986 to 2000 was estimated at 7.4 mgd (Todd Engineers, 2005).

The initial average hydraulic head in the Kalihi basal-water body was 40 ft. After forced draft started, the hydraulic head declined. The current average head is about 21 ft (Todd Engineers, 2005).

5.5.2. RAM2 Parameters

5.5.2.1. Mean Hydraulic Residence Time

The aquifer outflow is the difference between the rate of natural recharge and forced draft, or $Q = 12.0 - 7.4 = 4.6$ mgd, or 0.615×10^6 ft^3/d. The current average hydraulic head is

Table 10. Production Wells in Kalihi Basal-Water Body, 1986–2000

Well Name	Owner	Well No.	Number of Wells	GSE (ft)	Casing Depth (ft)	Well Depth (ft)	Elevation Below MSL (ft)
Kalihi Pump Station	BWS	1952	6	21	N/A	460	-439
			7	20	229	475	-455
			8	21	N/A	490	-469
			16	21	240	430	-409
			17	19	363	401	-382
			18	24	265	442	-418
			19	23	271	414	-391
			22	24	251	360	-336
Kapalama	BWS	2052	13	189	239	339	-150
			14	196	246	346	-150
Kamehameha School	Private	2051	1	549	604	705	-156
		2052	7	80	118	321	-241
		2052	11	90	129	334	-244
Jonathan Springs	Private	2052	12	31	50	151	-120
Dole Company	Private	1952	20	5	485	540	-535
			11	5	492	513	-508
			13	4	550	650	-646
			21	4	486	612	-608
Average							-370

NOTE: GSE = ground surface elevation, MSL = mean sea level, BWS = Board of Water Supply.

about 20.9 ft, and the area is 6.33 mi^2, or 1.77×10^8 ft^2. The present volume of freshwater can be calculated as $V = 41(h)(n_0)(A) = 41(20.9)(0.1)(1.77 \times 10^8) = 1.52 \times 10^{10}$ ft^3. Therefore, the mean hydraulic residence time of water particles in the Kalihi basal-water body under current conditions can be estimated as follows:

$$\tau = \frac{V}{Q} = \frac{1.52 \times 10^{10}}{0.615 \times 10^6} = 24,715 \text{ d}$$

5.5.2.2. Dispersion Coefficient

The effective dispersion coefficient of the Kalihi basal-water body is based on data taken at Jonathan Springs deep monitoring well and Kalihi Station deep monitoring well (Figure 24). It is determined using a curve-fitting process that compares the observed and calculated salinity profiles. The estimated time-of-water travel of the water particle and an effective porosity of 0.1 are used in the calculation. The time required for water particles to travel from the upper boundary of the basal-water body to Jonathan Springs deep monitoring well is 6,110 days and the time to Kalihi Station deep monitoring well is 7,280 days (Table 11).

Table 11. Water Particle Time-of-Water Travel to Deep Monitoring Wells in Kalihi Basal-Water Body

Parameter	Monitoring Well	
	Jonathan Springs	Kalihi Station Deep
High water head[a] (ft)	23	23
Low water head[a] (ft)	19.7	18.5
Hydraulic conductivity, K^a (ft/d)	1,000	1,000
Travel distance, L^b (ft)	14,200	18.100
Hydraulic gradient (ft)	2.3239E-04	2.4862E-04
Velocity, u (ft/d)	0.232	0.249
Porosity, P	0.10	0.10
True velocity, u' (ft/d)	2.32	2.49
Time-of-water travel, t_R (d)	6,110	7,280

[a]Data from Todd Engineers (2005).
[b]GIS-measured data.

The observed salinity profile for the Jonathan Springs deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 32. The estimated dispersion coefficient is 0.662 ft^2/d. The observed salinity profile for Kalihi Station deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 33. The estimated dispersion coefficient is 0.726 ft^2/d. The effective dispersion coefficient for the Kalihi basal-water body is taken to be the average of these two values, or $D_z = (0.662 + 0.726)/2 = 0.674$ ft^2/d.

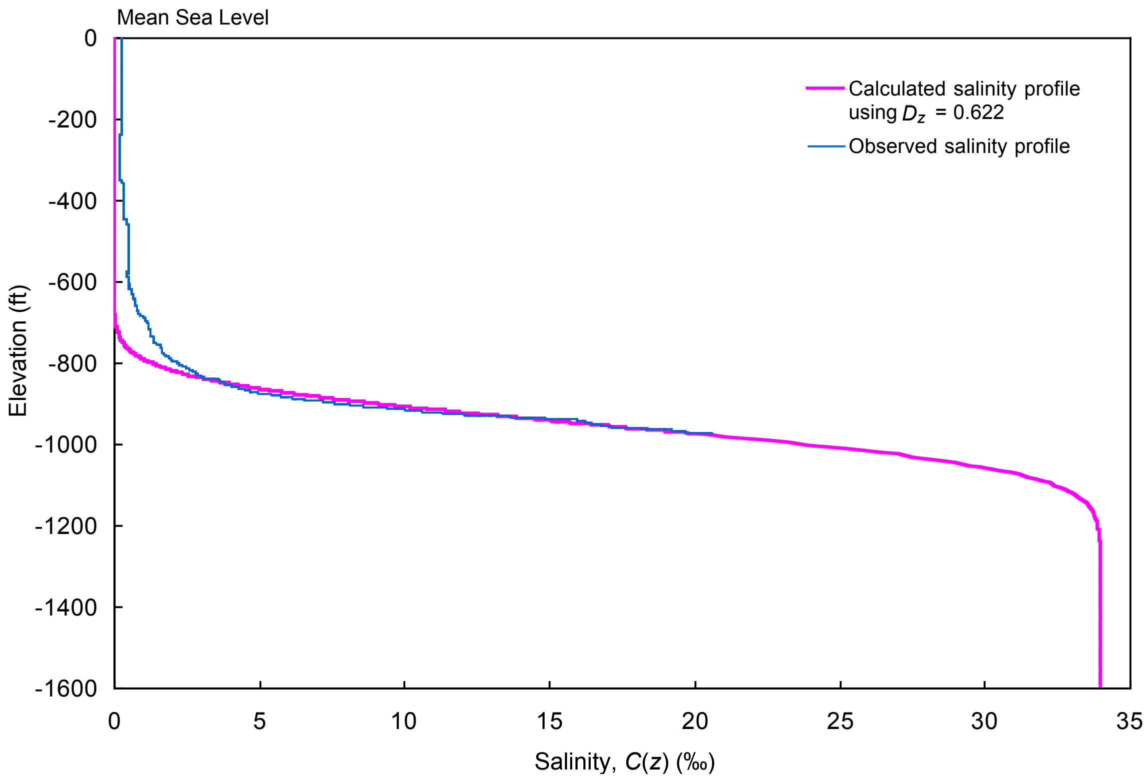


Figure 32. Estimation of dispersion coefficient by observed data fitting for Jonathan Springs deep monitoring well

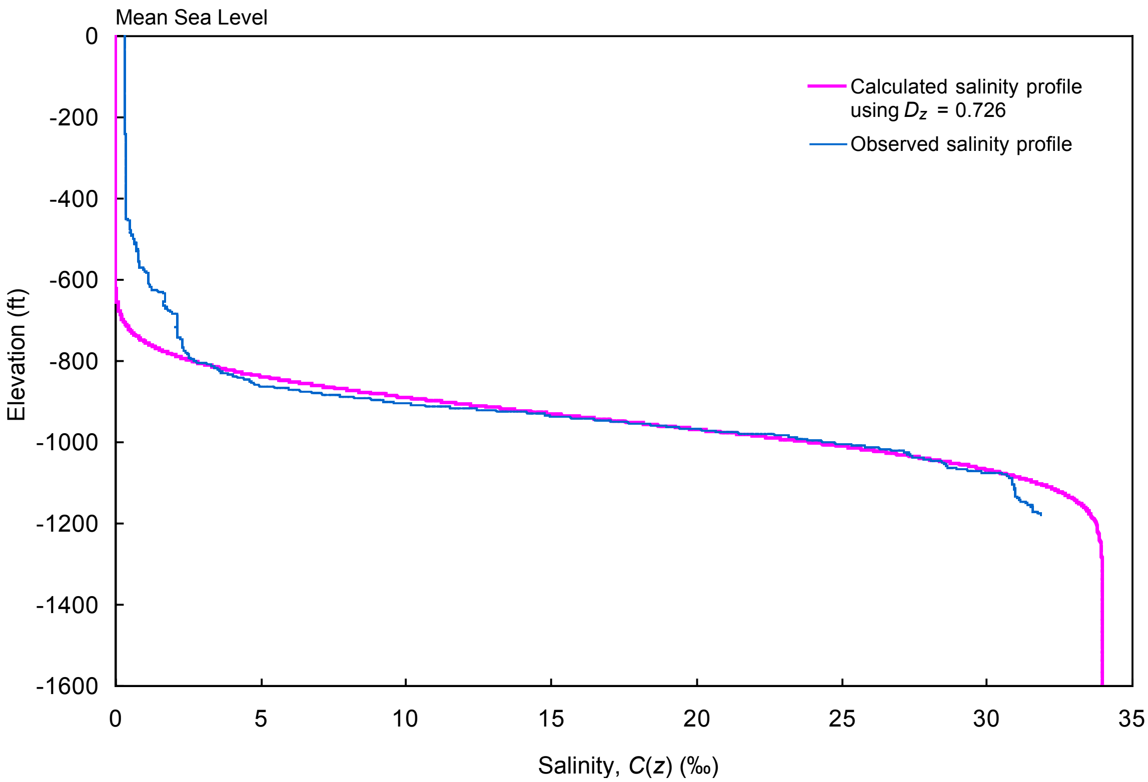


Figure 33. Estimation of dispersion coefficient by observed data fitting for Kalihi Station deep monitoring well

5.5.3. Sustainable Yield Determination

Determination of the minimum equilibrium hydraulic head for Kalihi basal-water body by transport submodeling is illustrated in Figure 34. Values of effective dispersion coefficient and mean hydraulic residence time are used to calculate the upper limit of the transition zone using Equation (4a). In this case, D_z = 0.674 ft^2/d and τ = 24,720 d, so ζ = 374 ft. To determine the effective well depth below mean sea level, the well depth and the upconing height are added (Equation (5)). The average depth of production wells in the Kalihi basal-water body is 370 ft below mean sea level, and the upconing height is 100 ft (Liu, 2006). Thus, η = 470 ft. Values for both the upper limit of the transition zone and the effective well depth below mean sea level are used to calculate the minimum equilibrium head (Equation (6)). The result is as follows: $h_e = (374 + 470)/40 = 21.10$ ft.

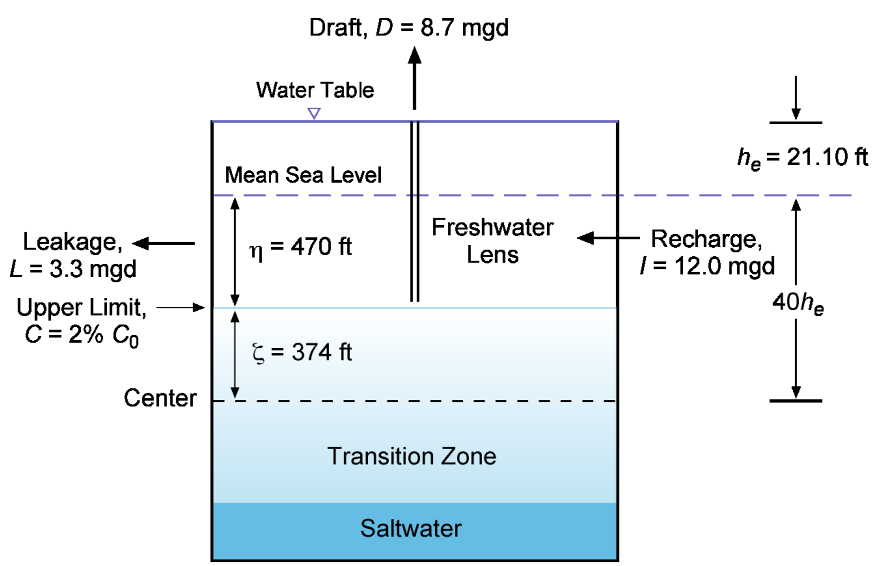


Figure 34. RAM2 transport submodeling for the determination of the equilibrium head of Kalihi basal-water body

Dividing the minimum equilibrium hydraulic head by the initial average water level, estimated at 40 ft (CWRM, 1990), gives the dimensionless variable of the minimum equilibrium hydraulic head. Thus, h_e/h_0 = 21.10/40 = 0.53. Based on the parabolic relationship between hydraulic head and forced draft derived by RAM, as illustrated in Figure 35, the dimensionless variable of forced draft is determined as D_s/I = 0.72.

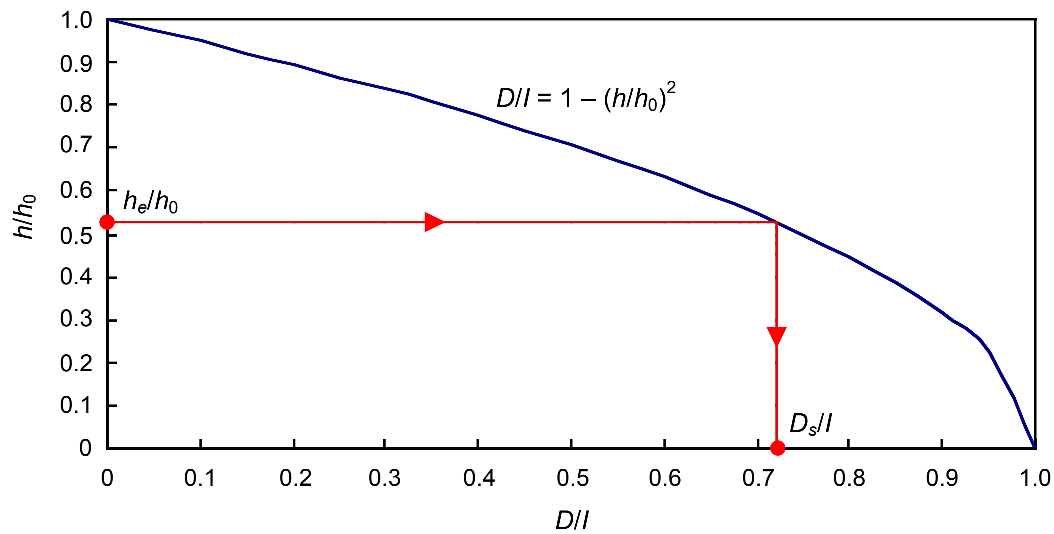


Figure 35. Determination of sustainable yield of Kalihi basal-water body by RAM2

The final step in determining the sustainable yield of a basal-water body using RAM2 is to multiply the rate of natural recharge by the dimensionless variable of draft. Thus, for Kalihi basal-water body the sustainable yield is calculated as $D_s = 12 \times 0.72 = 8.7$ mgd.

5.6. Estimation of the Sustainable Yield of Beretania Basal-Water Body

5.6.1. Aquifer Description

The Beretania basal-water body (State Code 30102) is designed groundwater area 2 by Honolulu Board of Water Supply. It is located between low-permeability valley-fill barriers underlying Nuuanu Stream and Manoa Stream (Figure 24). It covers an area of 8.62 mi^2.

Rainfall distribution over the Beretania basal-water body is illustrated by isohyetals in Figure 27. The average annual rainfall was estimated at 101 in. and the annual surface runoff and evaporation at 25 in. and 40 in., respectively. Thus the rate of natural recharge is about 36 in./yr, or 20.0 mgd.

Data on active pumping wells in the Beretania basal-water body and their respective depths below mean sea level are provided in Table 12. Total average draft for 1986 to 2000 was estimated at 16.3 mgd (Todd Engineers, 2005).

Table 12. Production Wells in Beretania Basal-Water Body, 1986–2000

Well Name	Owner	Well No.	Number of Wells	GSE (ft)	Casing Depth (ft)	Well Depth (ft)	Elevation Below MSL (ft)
Beretania Station	BWS	1851	12	21	499	580	-559
			13	22	499	616	-594
			24	20	487	616	-596
			25	17	495	617	-600
			31	20	489	600	-580
			32	20	489	600	-580
			33	14	479	533	-519
			34	14	484	636	-622
			35	15	488	566	-551
			67	18	484	619	-601
			74	19	498	598	-579
			75	18	523	625	-607
Wilder	BWS	1849	13	53	354	475	-422
			14	59	343	460	-401
			15	47	321	445	-398
			16	56	345	466	-410
Queen’s Hospital	Private	1851	54	34	356	460	-426
Punahou School	Private	1849	10	36	195	315	-279
Pacific Club	Private	1851	7	28	498	560	-532
Kawaiahao Church	Private	1851	73	13	709	777	-764
Average							-531

NOTE: GSE = ground surface elevation, MSL = mean sea level, BWS = Board of Water Supply.

Pre-development average hydraulic head in the Beretania basal-water body was 42 ft. Over the years, forced draft caused the hydraulic head to decline to the current average head of about 24 ft (Todd Engineers, 2005).

5.6.2. RAM2 Parameters

5.6.2.1. Mean Hydraulic Residence Time

The aquifer outflow is the difference between the rate of natural recharge and forced draft, or Q = 20.0 – 16.3 = 3.7 mgd. The current average hydraulic head is about 24.0 ft, and the area is 8.62 mi^2, or 2.40×10^8 ft^2. The present volume of freshwater can be calculated as $V = 41(h)(n_0)(A) = 41(24.0)(0.1)(2.40 \times 10^8) = 2.37 \times 10^{10}$ ft^3. Therefore, the mean hydraulic residence time of water particles in the Beretania basal-water body under current conditions can be estimated as follows:

$$\tau = \frac{V}{Q} = \frac{2.37 \times 10^{10}}{3.7(10^6 \times 0.13368)} = 47,879 \text{ d}$$

5.6.2.2. Dispersion Coefficient

The effective dispersion coefficient of the Beretania basal-water body is based on data taken at Beretania deep monitoring well and Punchbowl deep monitoring well (Figure 24). It is determined using a curve-fitting process that compares the observed and calculated salinity profiles. The estimated time-of-water travel of the water particle and an effective porosity of 0.1 are used in the calculation. The time required for water particles to travel from the upper boundary of the basal-water body to Beretania deep monitoring well is 16,239 days and the time to Punchbowl deep monitoring well is 15,931 days (Table 13).

Table 13. Water Particle Time-of-Water Travel to Deep Monitoring Wells in Beretania Basal-Water Body

Parameter	Monitoring Well	
	Beretania Deep	Punchbowl Deep
High water head (ft)	24.8	24.8
Low water head (ft)	22.2	23.9
Hydraulic conductivity, K (ft/d)	1,000	1,000
Travel distance, L (ft)	17,700	13,860
Hydraulic gradient (ft)	1.4689E-04	5.8000E-05
Velocity, u (ft/d)	0.147	0.058
Porosity, P	0.10	0.10
True velocity, u' (ft/d)	1.09	0.87
Time-of-water travel, t_R (d)	16,239	15,931

The observed salinity profile for Beretania deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 36. The estimated dispersion coefficient is 0.315 ft^2/d. The observed salinity profile for the Punchbowl deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 37. The estimated dispersion coefficient is 0.120 ft^2/d. The effective

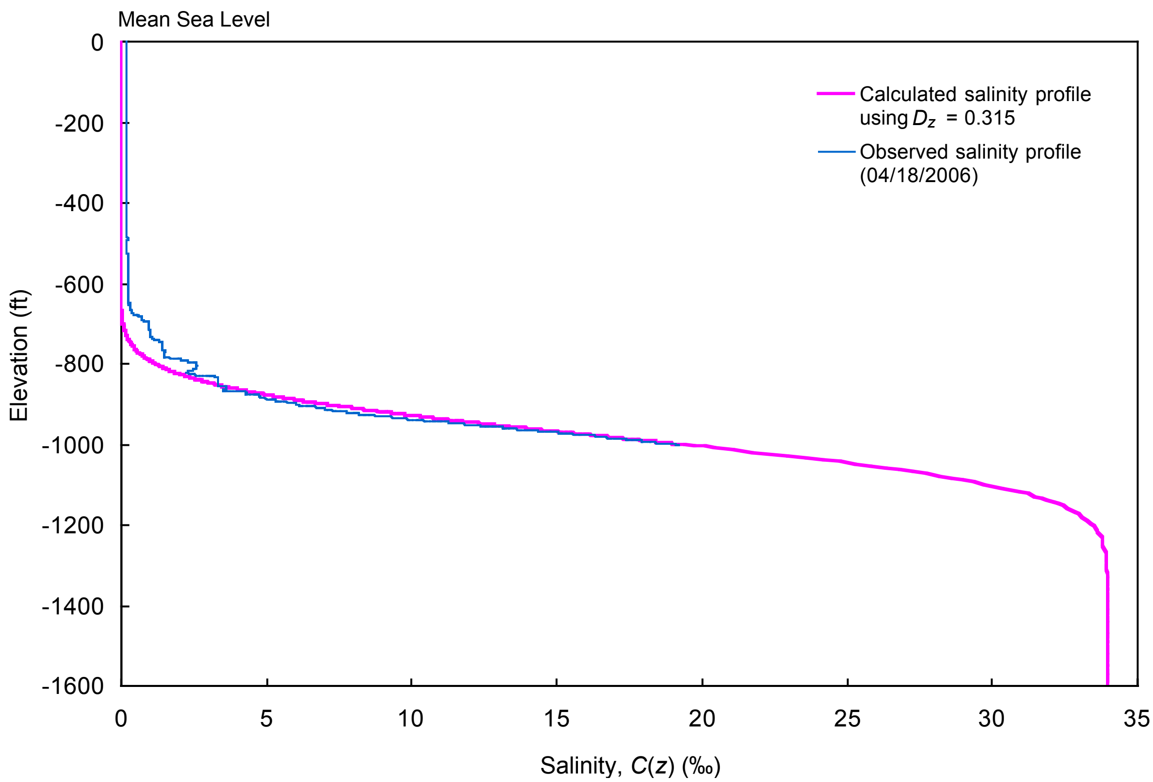


Figure 36. Estimation of dispersion coefficient by observed data fitting for Beretania deep monitoring well

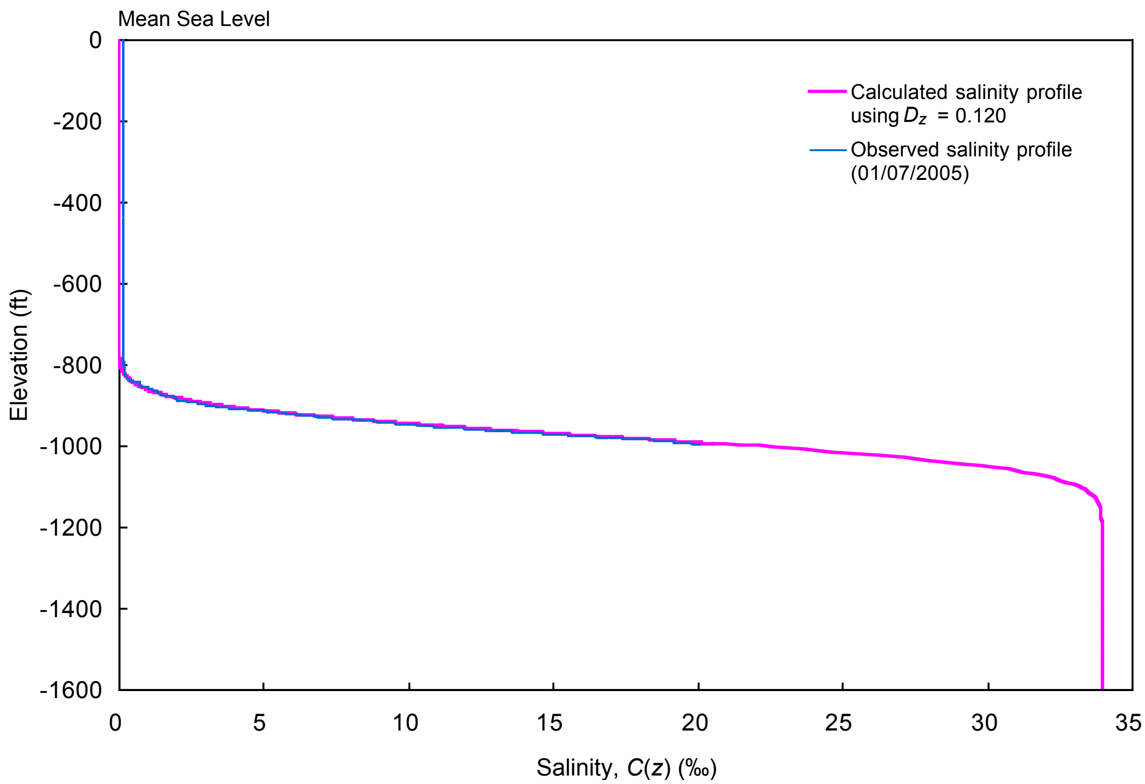


Figure 37. Estimation of dispersion coefficient by observed data fitting fort Punchbowl deep monitoring well

dispersion coefficient for the Beretania basal-water body is the average of these two values, or $D_z = (0.315 + 0.120)/2 = 0.218$ ft^2/d.

5.6.3. Sustainable Yield Determination

The minimum equilibrium hydraulic head calculated by the RAM2 transport submodel for Beretania basal-water body is illustrated in Figure 38. Values for effective dispersion coefficient and mean hydraulic residence time are used to calculate the upper limit of the transition zone using Equation (4a). In this case, $D_z = 0.218$ ft^2/d and $\tau = 47{,}879$ d, so $\zeta = 296$ ft. To determine the effective well depth below mean sea level, the well depth and the upconing height are added (Equation (5)). The average depth of production wells in the Beretania basal-water body is about 531 ft below mean sea level and the upconing height is about 100 ft (Liu, 2006). Thus, $\eta = 631$ ft. Values for both the upper limit of the transition zone and the effective well depth below mean sea level are used to calculate the minimum equilibrium head (Equation (6)). The result is as follows: $h_e = (296 + 631)/40 = 23.18$ ft.

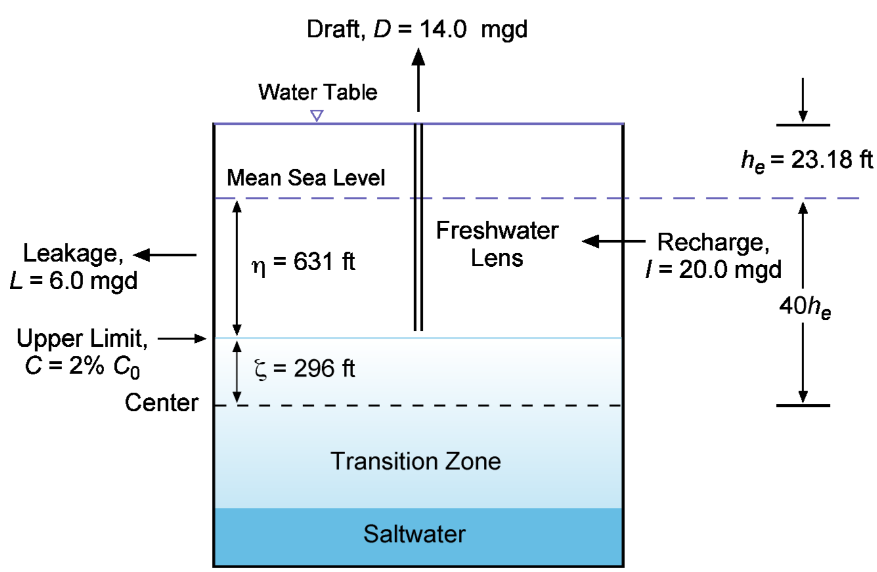


Figure 38. RAM2 transport submodeling for the determination of the equilibrium head of Beretania basal-water body

Dividing the minimum equilibrium hydraulic head by the initial average water level gives the dimensionless variable of the minimum equilibrium hydraulic head. Thus, h_e/h_0 =

23.18/42 = 0.55. Based on the parabolic relationship between hydraulic head and forced draft derived by RAM, as illustrated in Figure 39, the dimensionless variable of forced draft is determined as $D_s/I = 0.70$.

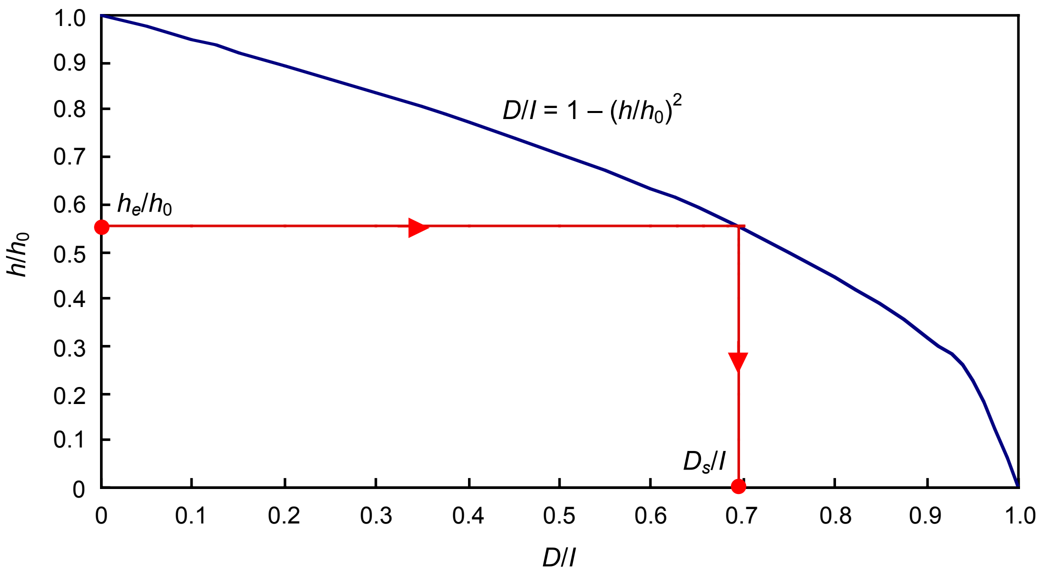


Figure 39. Determination of sustainable yield of Beretania basal-water body by RAM2

The final step in determining the sustainable yield of a basal-water body using RAM2 is to multiply the rate of natural recharge by the dimensionless variable of draft. Thus, for Beretania basal-water body the sustainable yield is calculated as $D_s = 20 \times 0.70 = 14.0$ mgd.

5.7. Estimation of the Sustainable Yield of Kaimuki Basal-Water Body

5.7.1. Aquifer Description

The Kaimuki basal-water body (State Code 30101) is designated groundwater area 1 by Honolulu Board of Water Supply. It is located between low-permeability valley-fill barriers underlying Manoa Stream and Palolo Valley (Figure 24). It covers an area of 4.43 mi^2 (CWRM, 1990).

Rainfall distribution over the Kaimuki basal-water body is illustrated by isohyetals in Figure 27. The average annual rainfall was estimated at 82 in. and the annual surface runoff and evaporation at 16 in. and 40 in., respectively. According to Todd Engineers (2005), the Kaimuki subbasin in Koolau basalt recharge is about 7.2 mgd and the alluvial recharge about 1.5 mgd. Based on these figures, the total rate of natural recharge is about 8.7 mgd for the Kaimuki basal-water body.

Data on active pumping wells in the Kaimuki basal-water body and their respective depths below mean sea level are provided in Table 14. Total average draft for 1986 to 2000 was estimated at 5.9 mgd (Todd Engineers, 2005).

Table 14. Production Wells in Kaimuki Basal-Water Body, 1986–2000

Well Name	Owner	Well No.	Number of Wells	GSE (ft)	Casing Depth (ft)	Well Depth (ft)	Elevation Below MSL (ft)
Kaimuki Station	BWS	1748	3	25	N/A	260	-235
			4	26	N/A	260	-234
			5	37	100	250	-213
			6	27	100	250	-223
			7	29	95	301	-272
			8	28	101	304	-276
			9	28	101	302	-274
			10	32	101	308	-276
Palolo	BWS	1847	1	387	420	570	-183
Sheraton	Private	1749	19	21	175	276	-255
Love's Bakery	Private	1749	18	16	255	269	-253
Average							-245

NOTE: GSE = ground surface elevation, MSL = mean sea level, BWS = Board of Water Supply.

The initial average hydraulic head in the Kaimuki basal-water body was 38 ft. Over the last 100 years, forced draft caused the hydraulic head to decline. The current average head is about 24 ft (Todd Engineers, 2005).

5.7.2. RAM2 Parameters

5.7.2.1. Mean Hydraulic Residence Time

The aquifer outflow is the difference between the rate of natural recharge and forced draft. The rate of natural recharge in the Kaimuki aquifer was estimated at 6 mgd by the Hawaii Water Resources Protection Plan (CWRM, 1990). Since the 1920s, the average rate of forced draft has been 5.9 mgd. At this rate, there has been no noticeable hydraulic head decline or chloride increase in pumped water. Therefore, the rate of natural recharge in the Kaimuki aquifer may be underestimated in the Hawaii Water Resources Protection Plan. In this study, the recharge rate of 8.7 mgd, as estimated by Todd Engineers (2005), is used. Using the natural recharge rate of 8.7 mgd and the forced draft rate of 5.9 mgd, the aquifer outflow is calculated as $Q = 8.7 - 5.9 = 2.8$ mgd, or 0.802×10^6 ft^3/d.

The current average hydraulic head is about 24 ft, and the aquifer area is 4.43 mi^2, or 1.24×10^8 ft^2. The present volume of freshwater can be calculated as $V = 41(h)(n_0)(A) = 41(24.0)(0.1)(1.24 \times 10^8) = 1.22 \times 10^{10}$ ft^3.

The mean hydraulic residence time of water particles in the Kaimuki basal-water body under current conditions can be estimated as follows:

$$\tau = \frac{V}{Q} = \frac{1.22 \times 10^{10}}{0.802 \times 10^6} = 15{,}212 \text{ d}$$

5.7.2.2. Dispersion Coefficient

Groundwater in the Kaimuki basal-water body flows in a mountain to ocean direction (Figure 26). Within the basal-water body, spatial variation of the hydraulic head is small. The simulated heads at three deep monitoring wells (i.e., Waahila Deep, Kaimuki High School Deep, and Kaimuki Station Deep), which are located far apart, are all about 26 ft (Figure 27) (Todd Engineers, 2005). With such a small variation of hydraulic head, it is difficult to calculate hydraulic gradient and flow velocity.

For the other three basal-water bodies in Honolulu aquifer (discussed previously in this report), the Darcy velocity of groundwater flow is determined by multiplying the hydraulic gradient times the hydraulic conductivity. The true velocities of groundwater flow are then calculated by dividing the Darcy velocity by the effective porosity. As shown in Tables 9, 11, and 13, the respective true groundwater flow velocities for the other three basal-water bodies range from 0.87 ft/d to 5.00 ft/d.

Due to the unavailability of reliable hydraulic gradient data, the true velocity of groundwater flow in the Kaimuki basal-water body cannot be calculated. The velocity is assumed to be 0.5 ft/d in this study. The respective time-of-water travel to the three deep monitoring wells can be calculated by dividing the travel distance (from the upper boundary of the aquifer to the well) by the true velocity. Thus, dividing 0.5 ft/d into 7,180 ft for Waahila Deep, into 13,860 ft for Kaimuki HS Deep, and into 17,700 ft for Kaimuki Station Deep gives time-of-water travel of 14,360 days, 27,720 days, and 35,400 days, respectively.

The effective dispersion coefficient of the Kaimuki basal-water body is determined using a curve-fitting process that compares the observed salinity profile at each monitoring well with its calculated salinity profile. The estimated time-of-water travel of the water particle and an effective porosity of 0.1 are used in the calculation.

The observed salinity profile for Kaimuki HS deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 40. The estimated dispersion coefficient is 0.343 ft^2/d. The observed salinity profile for Kaimuki deep monitoring well, along with the calculated profile that gives the minimum square error, is shown in Figure 41. The estimated dispersion coefficient is 0.594 ft^2/d. The observed salinity profile for Waahila deep monitoring well (located in Manoa Valley near the boundary between the Kaimuki and Beretania basal-water bodies), along with the calculated profile that gives the minimum square error, is shown in Figure 42. The estimated dispersion coefficient is 1.058 ft^2/d. The effective dispersion coefficient for the Kaimuki basal-water body is the average of these three values, or $D_z = (0.343 + 0.594 + 1.058)/3 = 0.665$ ft^2/d.

5.7.3. Sustainable Yield Determination

The minimum equilibrium hydraulic head calculated by the RAM2 transport submodel for Kaimuki basal-water body is illustrated in Figure 43. Values for effective dispersion coefficient and mean hydraulic residence time are used to calculate the upper limit of the transition zone using Equation (4a). In this case, $D_z = 0.665$ ft^2/d and $\tau = 15{,}212$ d, so $\zeta = 427$ ft. To determine the effective well depth below mean sea level, the well depth and the upconing height are added (Equation (5)). The average depth of production wells in the Kaimuki basal-water body is about 245 ft below mean sea level and the upconing height is about 100 ft (Liu, 2006). Thus, $\eta = 345$ ft. Values for both the upper limit of the transition zone and the effective well depth below mean sea level are used to calculate the minimum equilibrium head (Equation (6)). The result is as follows: $h_e = (427 + 345)/40 = 19.30$ ft.

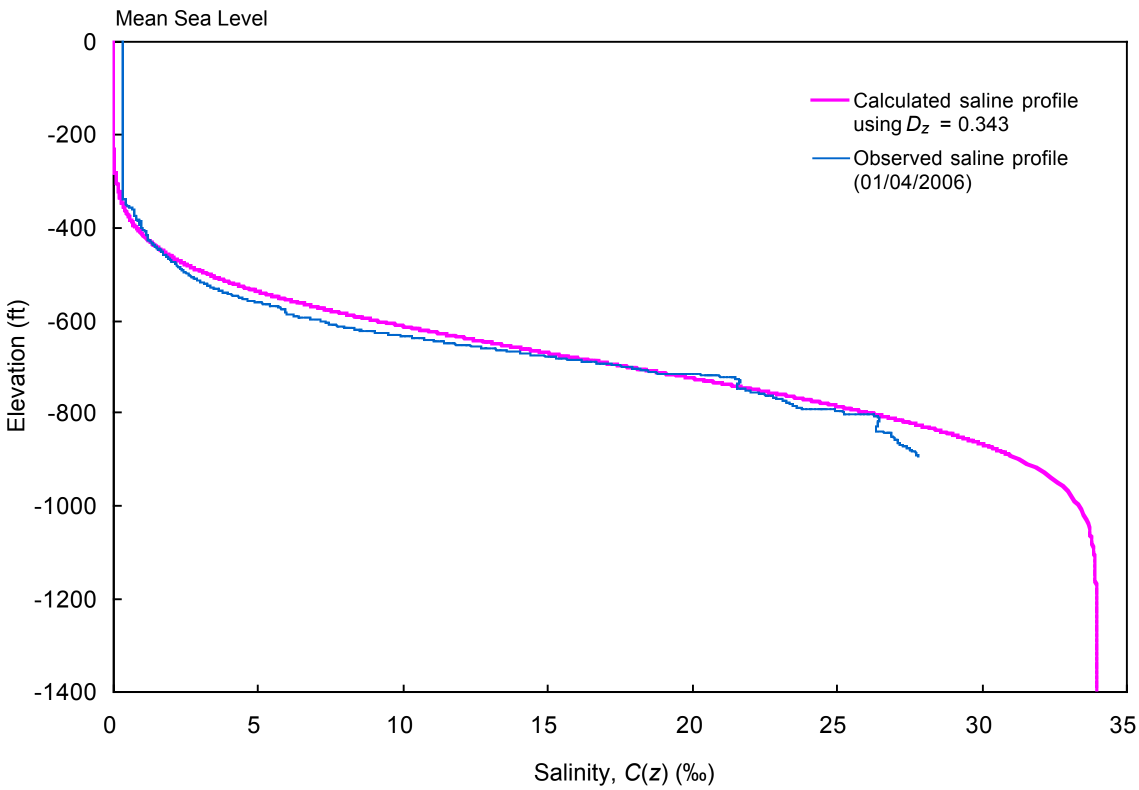


Figure 40. Estimation of dispersion coefficient by observed data fitting for Kaimuki High School deep monitoring well

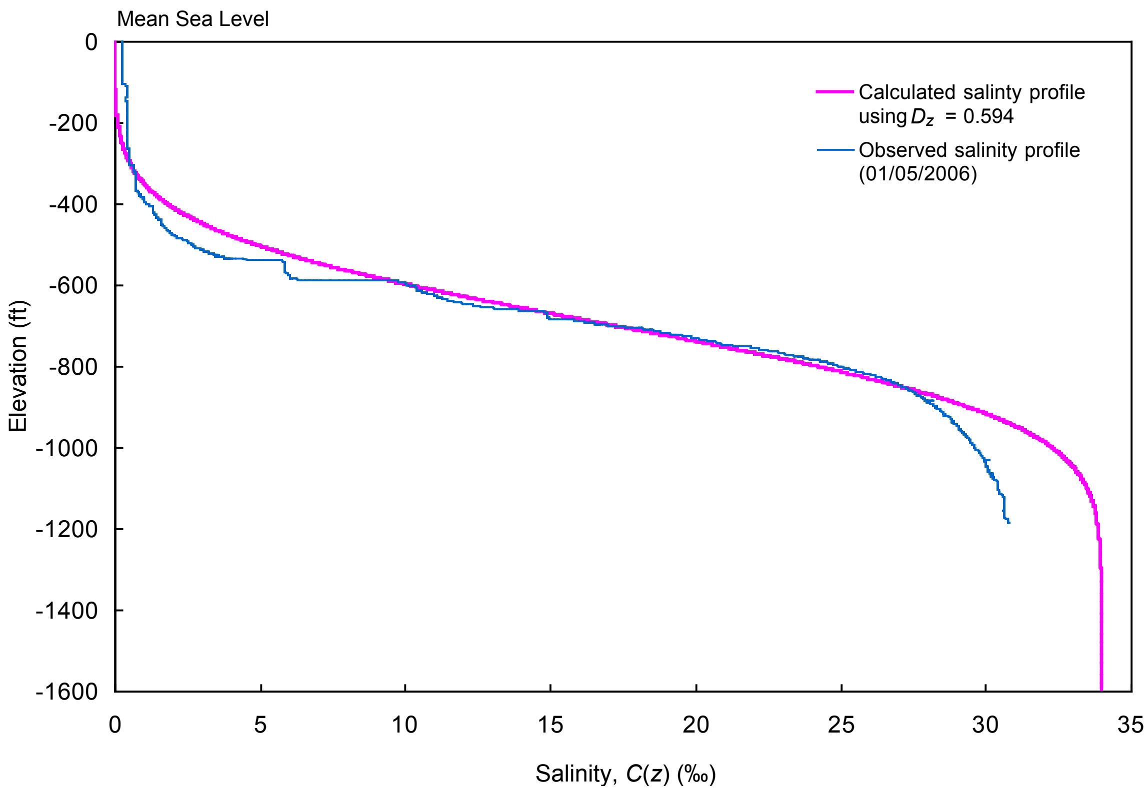


Figure 41. Estimation of dispersion coefficient by observed data fitting for Kaimuki deep monitoring well

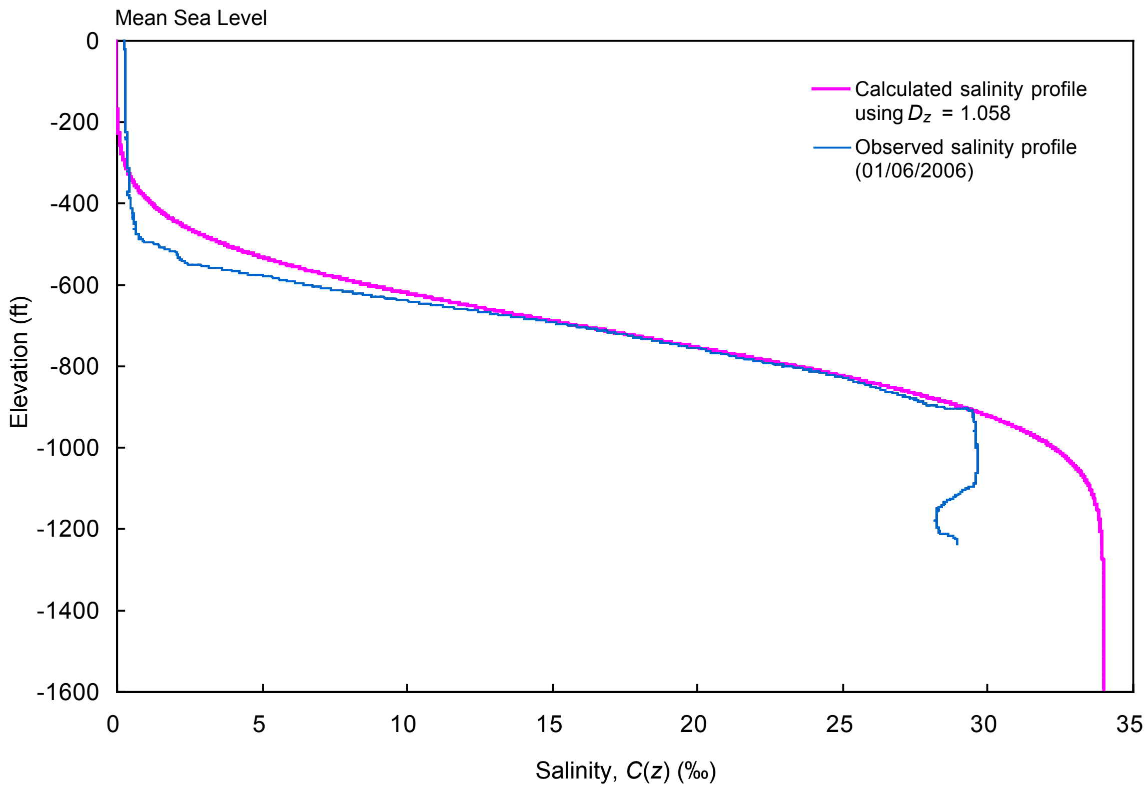


Figure. 42. Estimation of dispersion coefficient by observed data fitting for Waahila deep monitoring well

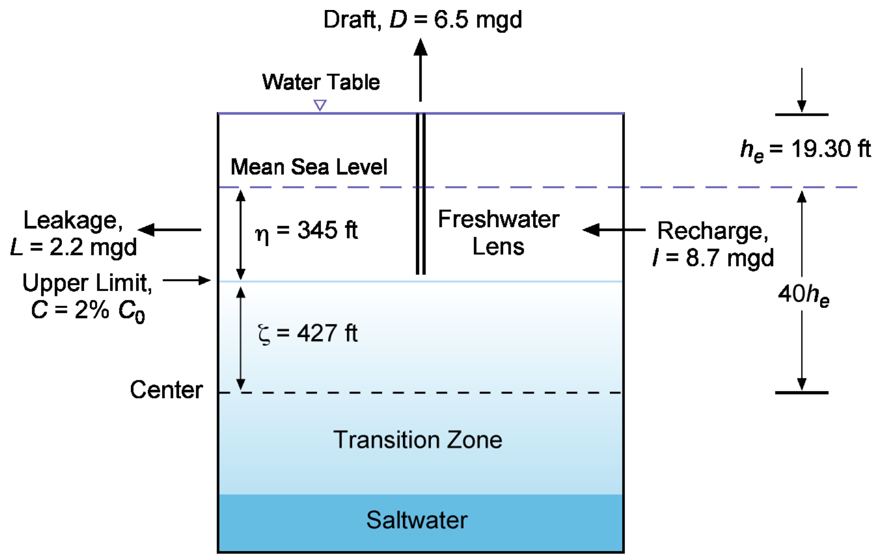


Figure 43. RAM2 transport submodeling for the determination of the equilibrium head of Kaimuki basal-water body

The dimensionless variable of the minimum equilibrium hydraulic head can be determined by dividing the minimum equilibrium hydraulic head by the initial average water level. Thus, $h_e/h_0 = 19.30/38 = 0.51$. Based on the parabolic relationship between hydraulic head and forced draft derived by RAM, as illustrated in Figure 44, the dimensionless variable of forced draft is determined as $D_s/I = 0.74$.

The final step in determining the sustainable yield of a basal-water body using RAM2 is to multiply the rate of natural recharge by the dimensionless variable of draft. Thus, for Kaimuki basal-water body the sustainable yield is calculated as $D_s = 8.7 \times 0.74 = 6.5$ mgd.

6. DISCUSSION

Estimating the sustainable yield of basal aquifers is a key element of groundwater management in Hawaii. Sustainable yield is defined as the amount of water that can normally be withdrawn from the aquifer at the maximum rate without unduly impairing the aquifer utility as a freshwater supply. The utility of a Hawaii basal aquifer may be impaired due to storage depletion and saltwater intrusion, both of which are consequences of overpumping.

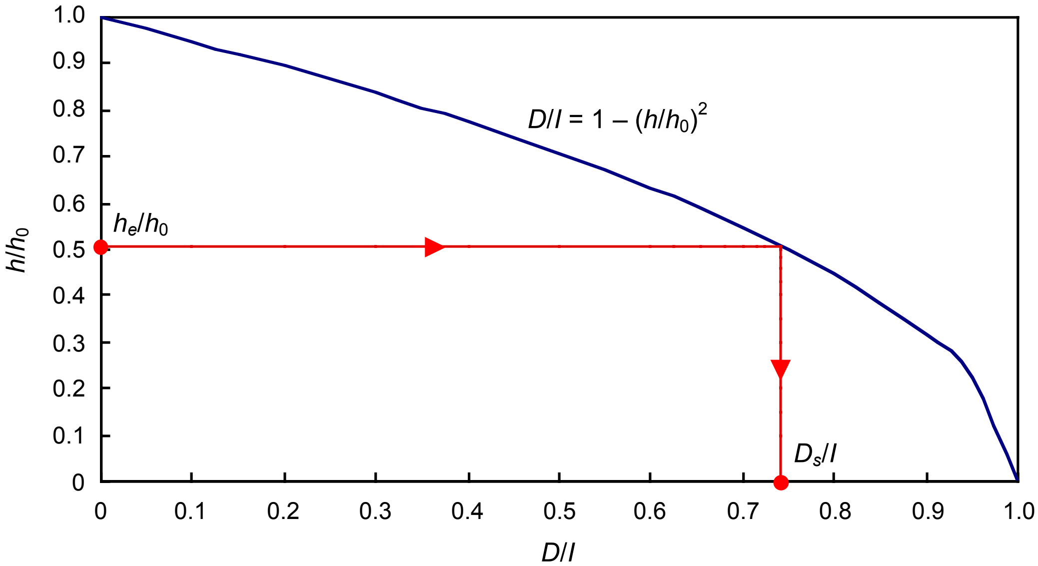


Figure 44. Determination of sustainable yield of Kaimuki basal-water body by RAM2

The storage of a basal aquifer is represented by its hydraulic head (Lau and Mink, 2006). Numerical modeling analysis of groundwater flow has been conducted to establish the relationship between the hydraulic head of a basal aquifer and the rate of pumping, or forced draft (Liu et al., 1983; Oki and Meyer, 2001). However, most of these models are too complicated for users to comprehend, and field data are not currently available for adequate parameter identification, calibration, and verification of these models.

A simple groundwater flow model called the robust analytical model (RAM) was developed by Mink (1980, 1981). By assuming a constant longitudinal water flow, the governing equation of RAM, or Equation (1), takes the form of a completely stirred tank reactor (CSTR). Equation (1) gives the variation of the spatially averaged hydraulic head of a basal aquifer in response to pumping stress. Two hydraulic parameters in Equation (1) are storage coefficient and initial hydraulic head. The initial hydraulic head of a basal aquifer denotes implicitly the aquifer's overall hydraulic conductivity and porosity.

The steady-state solution of RAM yields a simple parabolic relationship between the forced draft from a basal aquifer and its hydraulic head. The respective sustainable yields of major Hawaii basal aquifers estimated by RAM modeling are included in the state Water Resources Protection Plan, prepared by the Hawaii Commission on Water Resource Management (CWRM, 1990).

Because RAM simulates a basal aquifer as a CSTR, it does not address the effect of spatial hydraulic conductivity variation on groundwater flow. A USGS report (Oki and Meyer, 2001) suggests that one of the limitations of RAM is its inability to account for coastal caprock when estimating aquifer outflow. However, it should be noted that RAM was derived as an analytical tool for the estimation of sustainable yield and that its use for the study of spatial variability of groundwater flow is beyond the scope of its intended application.

As shown in Figure 2, the sustainable yield of a basal aquifer (D_s) can be readily determined when the value of the minimum equilibrium head (h_e) is known. Values of the minimum equilibrium head, essential for RAM modeling, were determined by using empirical relationships as shown in Table 2. Mink (1980) noticed at very beginning the empirical nature of the selection of minimum equilibrium heads. He suggested that an objective determination of the equilibrium head can only be made by combining the flow and salinity transport equations.

In a basal aquifer, the freshwater and underlying saltwater are separated by a transition zone, in which the salinity increases downward from 0% saltwater to 100% saltwater. In practice, the upper boundary of a transition zone is defined by the location where the salinity is 2% that of saltwater. The thickness of a transition zone depends on salt transport and mixing, a hydrodynamic dispersion process. It is noted that RAM was formulated on the assumption that a sharp interface exists between the freshwater in a basal aquifer and the underlying saltwater (Mink, 1980); thus, it cannot address the potential problem of saltwater intrusion.

RAM was modified to allow an investigation of the potential problem of saltwater intrusion by recognizing the existence of a transition zone (Liu, 2006). The modified RAM, or RAM2, consists of a flow submodel (which takes the form of RAM) and a transport submodel. The transport submodel simulates the expansion of the transition zone as the head declines. The modeling analysis by the transport submodel of RAM2 determines analytically the minimum equilibrium head of a basal aquifer. When the head is lower than the minimum equilibrium head, the salinity of water pumped out the aquifer would be higher than the acceptable level.

A natural field tracer method for estimating effective dispersion coefficients of Hawaii basal aquifers was developed. It uses salinity as a natural conservative tracer. By this method, the effective dispersion coefficient for the basal aquifer at each monitoring well is estimated by a curve-fitting process that compares the salinity profile observed at a deep monitoring well with that calculated using Equation (4b).

RAM2 was applied to re-evaluate the sustainable yield of six Hawaii basal-water bodies. A comparison of the sustainable yields estimated by RAM (CWRM, 1990) and RAM2 is shown in Table 15.

Table 15. Minimum Equilibrium Heads and Sustainable Yields Estimated by RAM and RAM2 for Six Hawaii Basal-Water Bodies

Aquifer	Minimum Equilibrium Head (ft)		Estimated Sustainable Yield (mgd)	
	RAM	RAM2	RAM	RAM2
Iao aquifer on Maui	14.0	15.85	20.0	19.0
Kualapuu aquifer on Molokai	7.5	8.62	7.0	5.0
Honolulu aquifer on Oahu				
Moanalua basal-water body	19.0	22.24	18.0	17.3
Kalihi basal-water body	20.0	21.10	9.0	8.7
Beretania basal-water body	21.0	23.18	15.0	14.0
Kaimuki basal-water body	19.0	19.30	5.0	6.5

7. CONCLUSIONS

RAM2 modeling for the re-evaluation of the sustainable yield of several Hawaii aquifers for which deep monitoring well data are available resulted in the following estimates: 19.0 mgd for Iao aquifer on Maui, 5.0 mgd for Kualapuu aquifer on Molokai, and 46.5 mgd for Honolulu aquifer on Oahu. The value for the Honolulu aquifer was determined by adding the sustainable yields of four basal-water bodies comprising the aquifer. The independent re-evaluation of these basal-water bodies resulted in the following sustainable yield estimation: 17.3 mgd for the Moanalua basal-water body, 8.7 mgd for the Kalihi basal-water body, 14.0 mgd for the Beretania basal-water body, and 6.5 mgd for the Kaimuki basal-water body.

The estimated sustainable yield of a Hawaii basal aquifer depends on the values of two RAM2 transport parameters, i.e., dispersion coefficient and mean hydraulic residence time. Mean residence time can be readily determined with known effective volume of the aquifer and outflow rate. Dispersion coefficients must be determined based on field measured salinity profiles at monitoring wells and time-of-water travel from the upper boundary of the aquifer to these wells. In this study, time-of-water travel to each deep monitoring well in Iao and Kualapuu aquifers is determined by modeling analysis using SHARP (Essaid, 1990). Further improvement of the estimation of dispersion coefficient can be achieved by installing more deep monitoring wells.

Another important factor which determines the accuracy of the sustainable yield estimation is natural recharge. Unfortunately, values of the natural recharge of Hawaii basal aquifers reported by different studies vary significantly. For instance, the rate suggested by the Hawaii Commission on Water Resource Management (1990) and used by this study to determine the sustainable yield of Iao aquifer is 28 mgd. The rate suggested by Shade (1997b) is 35 mgd, and that suggested by Engott (2006) is 44 mgd. Such discrepancies are mainly due to different assumptions concerning various components for a hydrologic balance analysis. Thus, more studies on individual hydrologic processes of watersheds in Hawaii are needed. The resulting data would not only lead to more accurate determination of sustainable yield of basal aquifers but also lead to better management planning of water quality and flood control.

The modeling techniques developed by this study are useful for the estimation of sustainable yield in Hawaii basal aquifers. However, because it is based on the conceptual formulation of a basal aquifer as a CSTR, it does not provide information on spatial variations of flow and transport in the basal aquifer. Comprehensive three-dimensional groundwater management models need to be developed for more detail investigations. Again, more deep monitoring wells will have to be installed to provide data for the calibration and verification of the groundwater models and for a fundamental study of the relationship between hydrodynamic dispersion in Hawaii basal aquifers and their basic hydrogeologic properties.

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APPENDIX

APPENDIX

Flow Simulation by SHARP Modeling and the Calculation of Time-of-Water Travel From the Upper Boundary of a Basal Aquifer to a Monitoring Well

SHARP is a quasi-three-dimensional, numerical finite-difference model for simulating freshwater and saltwater flow in a basal aquifer that is separated by a sharp interface. This model was developed and initially released by the U.S. Geological Survey in 1990 (Essaid, 1990). More recently, SHARP was used by the U.S. Geological Survey in a study on the effects of additional pumping on groundwater flow in central Oahu (Oki, 1998).

In our study, SHARP was applied to simulate groundwater flow in Iao aquifer on Maui and Kualapuu aquifer on Molokai. Velocity contours established by flow simulation allow satisfactory estimation of time-of-water travel from the upper boundary of the aquifer to the deep monitoring wells. The applicability of SHARP for flow simulation was tested by comparing simulated flow contours in the Pearl Harbor aquifer with observed data (Liu, 2006). The procedure and results of SHARP testing are included in this Appendix.

SHARP was applied to simulate the freshwater hydraulic head at steady state using the most recent pumping data and commonly accepted aquifer parameters as input data. In addition, it was used to estimate time-of-water travel, which is defined as the distance from the aquifer's upper boundary to the deep monitoring well divided by the groundwater velocity. Since the time-of-water travel is different for each monitoring well, depending on its location, it is important to calculate and prepare a water contour map for the basal aquifer. This map can be used to determine the path and velocity of groundwater movement.

The input data required for SHARP modeling includes (1) simulation parameters, (2) aquifer geohydrological parameters, and (3) pumping period. Simulation parameters and aquifer hydrogeological parameters were obtained from an earlier modeling exercise (Liu et al., 1983). Pumping data were obtained from files of "Pearl_Harbor_Production_Wells.xls" (provided by CWRM).

The modeling area and numerical segmentation of Pearl Harbor aquifer used for SHARP testing are similar to that used by Liu et al. (1983). The range of the latitude of the modeling area is from 21°16'00"N to 21°316'00"N and that of the longitude is from 157°50'00"W to 158°09'00"W. A 30 × 30 grid is used, with each cell enclosed by vertical and horizontal lines of 30" latitude and 38" longitude intervals (Figure A1). Each 30" latitude equals a length of 3,038.1 ft, and each 38" longitude equals a length of 3,592.6 ft. Therefore, the area of each cell is about 1.1 × 107 ft^2.

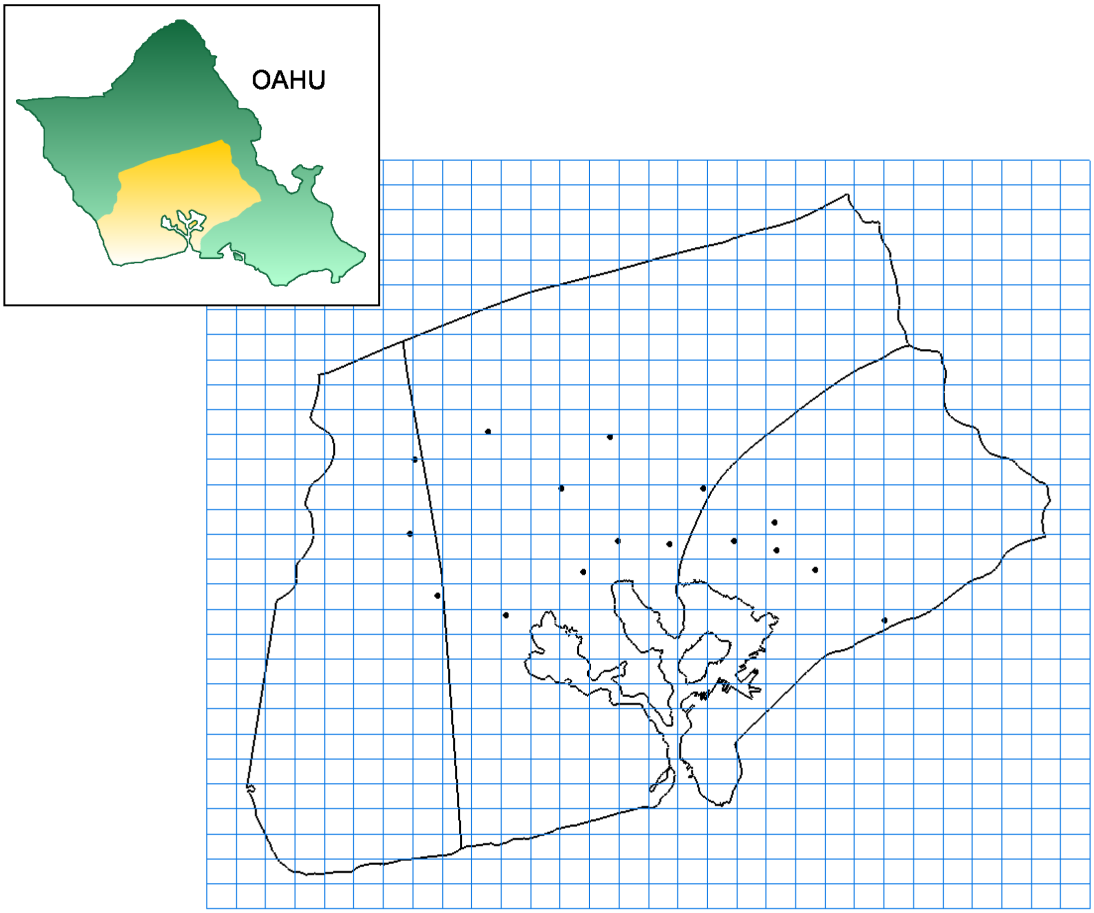

Figure A1. SHARP modeling of Pearl Harbor aquifer with a 30 × 30 grid network

A simulated map of the hydraulic head contour in Pearl Harbor aquifer and the observed head distribution during a synoptic hydraulic head survey conducted on May 15, 2003 are shown in Figure A2. The survey was a joint effort of the Hawaii Commission on Water Resource Management, Honolulu Board of Water Supply, and U.S. Geological Survey. The closeness of the simulated and observed head distributions indicates that SHARP is a satisfactory modeling tool for simulating groundwater flow in Hawaii basal aquifers.

(a)

NOTE: Contours in ft.

(b)

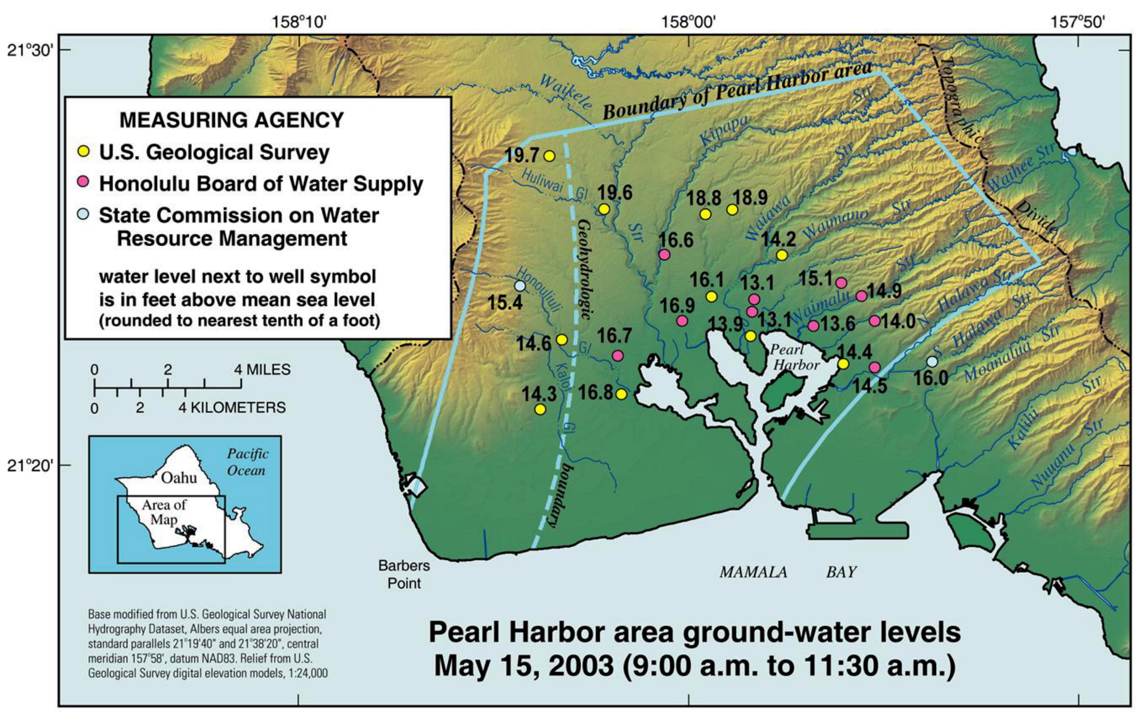

Figure A2. Hydraulic head distribution in the Pearl Harbor aquifer (a) as simulated by SHARP and (b) as observed on May 15, 2003