Three Dimensional Simulation of Seawater Intrusion in Heterogeneous Aquifers, with Application to the Coastal Aquifer of Israel

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ABSTRACT

This work presents a three-dimensional model of a portion of the upper section of the coastal aquifer in Israel, which suffers from seawater intrusion. The objective of the model is to serve as a management tool. The model was built by using the Finite Element framework, taking into account the development of a transition zone and the variation of fluid density within it. First, the model was run for long periods of time, without pumping, in order to create initial conditions of seawater intrusion (salt concentration distribution) prior to the exploitation of the aquifer. Then the model was used to analyze seawater intrusion induced by a pumping well in the vicinity of the coast (coastal collector). The model was able to adequately represent the behavior during periods of seawater intrusion, upcoming, and recovery, in response to well activation and shut-off.

INTRODUCTION

The coastal aquifer of Israel is one of the country's major sources of fresh water. It is intensively exploited through thousands of pumping wells. It serves also as a seasonal and long-term storage reservoir, in combination with artificial recharge through injection wells and infiltration ponds. This Pleistocene aquifer is heterogeneous, particularly in the vicinity of the coast, where clay layers divide it into 4 sub-aquifers that extend inland to some distance from the coast. It is commonly accepted that only the upper (A-)sub-aquifer is connected to the sea. As a result of its intensive exploitation, the aquifer has been experiencing seawater intrusion in many locations, forcing the shut-off of wells, particularly in the vicinity of the coast. In many locations restoration measures have been undertaken. In this work, seawater intrusion is analyzed as a case of density dependent flow and salt transport problem, with the help of a three-dimensional model. Three main scenarios are analyzed: 1) pristine conditions (no pumping; only recharge); 2) pumping above the seawater intrusion zone until the salinity reaches a threshold value (such as drinking water concentration criteria) and 3) recovery scenario in which pumping is stopped once the threshold salinity is reached.

MATHEMATICAL MODEL

Due to hydrodynamic dispersion, a transition zone develops in a coastal aquifer between freshwater and seawater. Across such zone, salt concentration varies from that of freshwater to that of (pure) seawater. In the model, freshwater and seawater are regarded
as a single liquid phase, referred as water, with a varying concentration of dissolved matter. We shall simulate seawater intrusion by developing a coupled density-dependent flow and salt transport mathematical model, transforming it into a numerical model, and solving the latter by an appropriate computer program.

**Governing Equations**

The density-dependent flow and salt transport equations that describe seawater intrusion in terms of reference head \( h_f = h_f(x, y, z, t) \) and normalized salt concentration \( c = c(x, y, z, t) \), take the form of two mass balance equations [Zhou, 1999, Bear, 1999]:

\[
\begin{align*}
S_0 \frac{\partial h_f}{\partial t} + \frac{\rho \phi \beta_c \partial c}{\rho_{f_0}} &= \frac{\partial}{\partial x_i} \left( K_{ij} \left( \frac{\partial h_f}{\partial x_j} + \beta_c \frac{\partial c}{\partial x_j} \right) \right) + \frac{(\rho^* Q_R - \rho Q_p)}{\rho_{f_0}}, \\
(l - \beta_c) \frac{Dc}{Dt} &= \frac{\partial}{\partial x_j} \left( \frac{\phi D_{ij}}{\rho_r} \frac{\partial c}{\partial x_j} \right) + Q_R \left( c^* - \frac{\rho^*}{\rho} \right),
\end{align*}
\]

with the rate of pumping and injection through wells represented, symbolically, by

\[
\begin{align*}
\rho Q_p &= \sum_m \rho_m Q_{pm}(x_m, t) \delta(x - x_m), \\
c Q_p &= \sum_m c_m Q_{pm}(x_m, t) \delta(x - x_m), \\
\rho^* Q_R &= \sum_n \rho^*_n Q_{Rn}(x_n, t) \delta(x - x_n), \\
c^* Q_R &= \sum_n c^*_n Q_{Rn}(x_n, t) \delta(x - x_n),
\end{align*}
\]

where \( x_i \) is the \( i \)th component of the Cartesian coordinates, \( x \) is ane position vector, \( h_f = p / \rho_{f_0} \) is the *reference freshwater head*, \( \rho \) is the fluid’s variable density, \( \rho_{f_0} \) is a *reference freshwater density*, \( \rho_r = (\rho / \rho_{f_0}) \) is the fluid’s relative density, \( p \) is the pressure, \( g \) is the gravity acceleration, \( z \) is the vertical component of \( x \), \( \phi \) is the porosity, \( \beta_c = (1/\rho) \partial \rho / \partial c \) is a coefficient that introduces the effect of concentration change on the fluid density, \( \beta_c^* = (1/\rho_{f_0}) \partial \rho / \partial c \) is the linear density coefficient, related to the reference density, \( K_{ij} = \rho g k_{ij} / \mu \) is the hydraulic conductivity tensor, \( k_{ij} \) is the permeability tensor, \( \mu \) is the dynamic viscosity, \( S_0 = \rho g (\phi \beta_p + \alpha) \) is the specific storativity of the porous medium, \( \beta_p = (1/\rho) \partial \rho / \partial p \) denotes the coefficient of water compressibility, \( \alpha = -(1/\rho) \partial \rho / \partial p \) denotes the coefficient of solid matrix compressibility (with \( \alpha = 0 \) for the incompressible material considered here), \( c = (C - C_{f_0}) / (C_s - C_{f_0}) \) is the normalized salt concentration, \( C \) is the salt concentration (mass of salt per unit volume of salt solution), and \( C_{f_0} \) and \( C_s \) are the salt concentrations
of pure freshwater and of pure seawater, respectively. $Q_{in}$, $\rho_n$, and $c_n$ are the injection rate, density, and salt concentration of water injected through a well at point $x_n$, respectively, $Q_{pm}$, $\rho_m$ and $c_m$ are the pumping rate, density, and salt concentration of water pumped through a well at point $x_m$, and $D_{ij}$ is the coefficient of hydrodynamic dispersion. Note that the hydraulic conductivity, $K_{ij}$, and the specific storativity, $S_0$, depend on salt concentration, and as such may vary in both space and time.

The salt transport equation, (2), is written in the Eulerian-Lagrangian framework, where $\frac{Dc}{Dt} \equiv \frac{\partial c}{\partial t} + V_i \frac{\partial c}{\partial x_i}$ is the material derivative, and $V_i$ is the $i$th component of the fluid’s velocity vector (relative to the solid matrix).

The generalized Darcy’s law can be written

$$q_i = \phi V_i = -\frac{K_{ij}^f}{\mu_r} \left( \frac{\partial h_f}{\partial x_j} + \beta_i^c \frac{\partial c}{\partial x_j} \right), \quad (3)$$

where $q_i$ is the $i$th component of the specific discharge vector $q$, $K_{ij}^f = \rho_f g k_{ij} / \mu_{f0}$ is the reference hydraulic conductivity, $\mu_{f0}$ is the reference freshwater dynamic viscosity, and $\mu_r = \mu / \mu_{f0}$ is the relative dynamic viscosity.

The coefficient of hydrodynamic dispersion, $D_{ij}$, is the sum of the coefficients of mechanical dispersion and of molecular diffusion in a porous medium, $D_m$. For an anisotropic porous medium [Bear, 1979; Bear and Verruijt, 1990]:

$$D_{ij} = a_{ijkm} \frac{V_i V_m}{V} + D_m \delta_{ij}, \quad (4)$$

where $a_{ijkm}$ is a typical component of the (fourth rank) dispersivity tensor, $V$ is the magnitude of the velocity, and $\delta_{ij}$ denotes the Kronecker delta. In an isotropic porous medium with $x_i$ serving as principal directions,

$$D_{ij} = a_L V \delta_{ij} + (a_L - a_T) \frac{V_i V_j}{V} + D_m \delta_{ij}, \quad (5)$$

where $a_L$ and $a_T$ are the longitudinal and transversal dispersivities, respectively.

**Constitutive Equations**

A typical relationship between fluid density, pressure, and salt concentration can be expressed in the form:

$$\rho = \rho_{f0} \exp \left[ \beta_p (p - p_0) + \beta_c c \right], \quad (6)$$

where $p_0$ is a reference pressure value. For seawater intrusion, we may employ the following linear expressions for variable density and dynamic viscosity:
\[ \rho = \rho_f \left(1 + \beta \epsilon \right), \]  
(7)

\[ \mu = \mu_f \left(1 + \beta \mu \right), \]  
(8)

where \( \beta \mu \) is a coefficient that introduces the effect of concentration changes on the fluid's dynamic viscosity.

**Initial and Boundary Conditions**

The initial conditions for the density-dependent flow and salt transport equations within the considered phreatic aquifer domain are

\[ h_f(x,0) = h_{f0}(x), \]  
(9a)

\[ c(x,0) = c_0(x), \]  
(9b)

where \( h_{f0}(x) \) and \( c_0(x) \) are known distributions of the reference head and the salt concentration, respectively.

**Fig. 1.** A typical cross-section of a phreatic coastal aquifer.

The conditions for density-dependent flow on the boundary segments are (Fig. 1):

\[ q_n = 0, \]  
\[ h_f = h_f^p, \text{ or } q_n = q_n^p, \]  
\[ \rho q_n = \left( \rho_N N - (\phi \rho - \theta \mu_0 \rho_N) \frac{\partial h_f}{\partial t} \right) n_z, \]  
\[ h_f = \xi_{DE}, \]  
\[ h_f = \beta \epsilon H_{\text{sea}}, \]  
\[ h_f = \beta \epsilon H_{\text{FA}}, \]

\begin{align*}
\text{on AB (im pervious bottom)} & \text{ on BC (land - side lateral boundary)} \\
\text{on CD (phreatic surface)} & \text{ on DE (seepage face)} \\
\text{on EF (sea bottom)} & \text{ on FA (sea - side laterla boundary)}
\end{align*}

(10)
where \( h_i^p \) is a prescribed (reference) head on a Dirichlet-type boundary, \( q_n^c \) is a prescribed fluid flux on a second-type boundary, \( n \) is the unit outward normal vector with components \( n_x, n_y, n_z \) in three dimensions, \( N \) denotes the replenishment, \( \rho_N \) is the density of \( N \), \( \theta_w \) is the irreducible water content assumed to prevail in the unsaturated zone above the phreatic surface, \( \xi_{DE} \) is the elevation of a point on the seepage face, \( H_{sea} \) and \( H_{FA} \) are the depths of seawater at a point on the sea bottom and on the sea-side lateral boundary, respectively.

The salt transport equation requires the following boundary conditions:

\[
\begin{align*}
q_n^d &\equiv -\phi D_{ij} \frac{\partial c}{\partial x_j} n_i = 0, & \text{on AB and DE} \\
c &= 0, & \text{on BC} \\
q_n^c &= \left( \frac{\rho_N - \theta}{\rho} \right) \left( N + \theta_w \frac{\partial h_i^p}{\partial t} \right) n_z, & \text{on CD} \\
c &= c_s = 1.0, & \text{on FA} \\
\end{align*}
\]

where \( q_n^d \) is the hydrodynamic dispersive flux normal to the boundary, \( c_s \) is the normalized salt concentration of pure seawater, and \( c_N \) is the salt concentration of the replenishment.

The boundary condition for transport on the sea bottom EF depends on the flow direction. An inflow (i.e., into the aquifer) and an outflow portion may exist along EF that are separated by a point with zero-normal fluid flux (point \( M \) in Fig. 1). For the inflow portion, we usually employ either a third-type condition

\[
q_n^d = (c_s - c) q_n^c, \quad (12a)
\]

or a Dirichlet condition

\[
c = c_s. \quad (12b)
\]

For the outflow portion, however, a third type condition in (12a) is never valid, since a Cauchy-type condition is applicable only for the inflow boundary; otherwise a contradiction will occur. To supply the required boundary condition, the assumption is often made that the fluid concentrations are identical on both sides of this boundary. Together with the fact that fluid fluxes are also identical, the condition degenerates into a second-type condition:

\[
q_n^d \equiv \phi D_{ij} \frac{\partial c}{\partial x_j} n_i = 0. \quad (12c)
\]
NUMERICAL SCHEME

A computer program, FEASi-SWIT, was developed in C++ for the simulation of seawater intrusion in three-dimensional groundwater systems, by solving the coupled density-dependent flow and salt transport equations. This code is a category of the code family, called FEASi (Finite Element Aquifer Simulation) that was verified in a number of benchmark cases [Zhou, 1999; Bensabat et al., 2000]. In three-dimensions, vertical triangular prism meshes are used to represent complex geometry of solution domains of interest. Corresponding bilinear basis functions are used for each prism element.

The Picard method is employed to linearize the nonlinear flow and transport equations in a sequential procedure. First, we solve the density-dependent flow equation, using the Galerkin finite element method (FEM), for the reference head field, given a salt concentration field in the previous iteration. Then we solve for the velocity field by performing the same FEM procedure on Darcy’s law. Finally, we solve the advective-dispersive transport equation by the FEM procedure within the Eulerian-Lagrangian framework.

We use an element-averaged fluid density for the body-force term in the FEM formulation of Darcy’s law in order to avoid a spurious velocity field in the vertical direction. This spurious velocity results from the inconsistency between the approximation of head gradient and fluid density (salt concentration) [Voss and Souza, 1987; Herbert et al., 1988]. In the FEASi-SWIT code, bilinear basis functions are used for both the reference head and the salt concentration in a three-dimensional system.

The advective-dispersive salt transport equation is solved in the Eulerian-Lagrangian framework. The Lagrangian concentration is derived by the adaptive pathline-based particle tracking algorithm presented by Bensabat et al. [2000]. We then solve the remaining dispersion equation using the Galerkin FEM. In the particle tracking algorithm, the tracking process is split along element boundaries on an inter-element basis. The sub-process within an element may be further refined along the particle’s path by subdividing the travel time within this element into a number of travel time increments. Whether this in-element pathline-based refinement is needed or not depends upon the complexity of the local velocity field. Tracking errors are controlled by practical criteria related to the rate of variation in the particle’s velocity (in magnitude and direction). A bilinear spatial and temporal interpolation of particle velocity is used to avoid the error introduced by the stepwise temporal approximation used in most existing models. The efficiency of the particle tracking is improved by adapting the tracking time in each tracking step. This adaptation is based on two particle velocity indices and on their corresponding given criteria. The tracking process in regions where the velocity varies significantly is split into more tracking steps than in regions with smooth velocity variations. This adaptive particle-tracking algorithm performs very well in a complex flow field in a density-dependent flow and transport problem [Zhou, 1999].

For each step in the above sequential procedure, we obtain a symmetric and positive definite linear algebraic systems for flow, velocity, or salt transport. A very efficient conjugate gradient solver, preconditioned by an incomplete Cholesky factorization, is employed for solving the resulting algebraic systems [Ajiz and Jennings, 1984].
APPLICATION TO A PHREATIC COASTAL AQUIFER

Problem Statement

The problem serves to demonstrate the application of the FEASi-SWIT code to a field with groundwater flow conditions similar to those found in the upper section of the Coastal Aquifer of Israel. The purpose of this problem is to investigate three situations: (1) seawater intrusion in the absence of groundwater extraction, (2) interface upconing to a well pumping above the transition zone, and (3) the decay of the upconed seawater mound after pumping stops, when the salinity of the pumped water exceeds a specified criterion (say, 2%).

The considered problem is described schematically in Fig. 2. The domain is simplified to a rectangular prism. It is 1000m long, 500m wide, and 105m thick. The aquifer domain is recharged by a constant freshwater influx on the left-hand-side boundary. On the other side, it is exposed to a body of seawater. The bottom boundary is impervious to flow and transport. The aquifer is recharged from above by natural replenishment through the phreatic surface. The front and back planes ($y = 0m$ and $y = 500m$) are impervious. The condition for transport on the seawater side boundary is specified to be a constant seawater concentration ($c = 1.0$) at the lower influx portion, and zero gradient of salt concentration normal to this boundary at the upper outflow portion (point M Fig. 1).

Fig. 2. Domain and boundary conditions in the application to a phreatic coastal aquifer.

Initially, the aquifer contains only freshwater. The transient process of seawater intrusion (Case A) is simulated until a steady-state solution is obtained. Then, we introduce a pumping well located above the transition zone at $x = 500m, y = 0m$. It is centered at $z = -10m$ with a 10m screen. The well pumps out 30% of the total influx of freshwater. After the salinity in the pumped water exceeds 2% of seawater, we shut off
the well, and the upconed saltwater mound undergoes decay (Case C). We also investigate the steady-state profiles of seawater intrusion and interface upconing under well pumpage (Case B).

**Numerical analysis**

We choose the set of model parameters shown in Table 1, based on flow and transport conditions in the Coastal Aquifer of Israel. The computational mesh consists of 16800 triangular prism elements. Uniform spacing of 50m, 25m, and 5m is employed in the \( x \), \( y \), and \( z \) directions, respectively. This mesh is fixed for any time step in the course of the simulation. The moving phreatic surface and the dynamic computational mesh is handled automatically by an active mesh in the model [Zhou, 1999].

**Table 1.** Model parameters for the upconing problem in a phreatic coastal aquifer.

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>Porosity</td>
<td>( \phi = 0.25 )</td>
</tr>
<tr>
<td>Reference horizontal conductivity</td>
<td>( K_{xx}^f = K_{yy}^f = 20m/d )</td>
</tr>
<tr>
<td>Reference vertical conductivity</td>
<td>( K_{zz}^f = 2m/d )</td>
</tr>
<tr>
<td>Natural replenishment</td>
<td>( N = 0.15m/)year</td>
</tr>
<tr>
<td>Lateral freshwater discharge</td>
<td>( q_0 = 0.0215m/d )</td>
</tr>
<tr>
<td>Pumping rate</td>
<td>( Q_w = 392.2m^3/d )</td>
</tr>
<tr>
<td>Density of pure freshwater</td>
<td>( \rho_{f0} = 1000kg/m^3 )</td>
</tr>
<tr>
<td>Density of pure seawater</td>
<td>( \rho_s = 1025kg/m^3 )</td>
</tr>
<tr>
<td>Density coefficient</td>
<td>( \beta_c = 0.025 )</td>
</tr>
<tr>
<td>Longitudinal dispersivity</td>
<td>( \alpha_L = 10m )</td>
</tr>
<tr>
<td>Transversal dispersivity</td>
<td>( \alpha_T = 1m )</td>
</tr>
<tr>
<td>Molecular diffusivity</td>
<td>( D_m = 0m^2/d )</td>
</tr>
</tbody>
</table>

**Results and discussion**

In Case A, the steady-state solution is obtained after 30 years. A longer simulation shows that no significant changes are observed. Fig. 3 shows the steady-state distribution of salt concentration, as well as the Ghyben-Herzberg sharp interface. The latter is derived from the computed position of the phreatic surface, using the well-known Ghyben-Herzberg relation [Bear, 1979; Bear and Verruijt, 1990]. The position of this interface is much further inland than the 0.5 isochlor in the transition simulation. Most of the 0.02 isochlor is located higher than this sharp interface. This feature makes pumping above the transition zone more difficult. The phreatic surface is at \( z = 2.419m \) on the freshwater side boundary, and \( z = 0m \) on the seaward side. Here, we do not simulate a seepage face, assuming that it is very small. Seawater enters the aquifer through the lower
portion of the seaward side boundary \((z \leq -60m)\). It then mixes with freshwater flowing from the opposite direction and exits at the upper outflow portion of this boundary. The total seawater influx entering the aquifer is 23.6\% of the total freshwater influx by recharge. This quantity decreases with the decrease in the dispersivity as the process of seawater intrusion proceeds. However, the velocity within the seawater wedge is not negligible. The recirculating flow cell, in the shape of a triangular prism, is about 650m long along the bottom boundary, and 60m high at the seaward side one. At the front of landward seawater flow, isochlors are more concentrated than elsewhere. Note that all isochlors are perpendicular, or tend to be perpendicular, to the bottom, while the sharp interface is tangential to the bottom boundary.

![Diagram of seawater intrusion](image)

**Fig. 3.** The steady-state salt concentration field in Case A.

Fig. 4 shows the salt concentration profiles in the plane of \(y = 0m\), in which the well is located, in the entire process of interface upconing. This process covers the transition from the steady-state solution in Case A to a new steady-state one under well pumpage in Case B. The latter is obtained at 30 years. The salt cone is initialized on the downstream side of the pumping well. Under the pumping well, it continuously rises and moves towards the well from the downstream direction. For example, the apex of the cone of the 0.1 isochlor at 2 years is located at \(x = 600m\), and \(z = -43m\). The regional flow to the sea also pushes the saltwater cone downstream. The “cone-like” isochlors occur only for low salt concentrations (e.g., \(c = 0.02\), or 0.1). However, the isochlors of higher concentration also globally rise towards the well at a distance from their corresponding position at initial time. At the same time, seawater intrudes further landward. For example, the 0.5 isochlor at the bottom moves from \(x = 420m\) at the initial time in Fig 4a to \(x = 290m\) at the new steady state in Fig. 4f. After a period of pumping, the well is contaminated by seawater, as illustrated in Fig. 5. The salinity of the pumped water reaches 2\% at 8.27 years, and is about 4.24\% at the final steady-state.
Fig 4. The transient process of interface upconing under well pumping in Case B.

Fig. 5. Breakthrough curves of the salinity in the water pumped from the well in Cases B and C. Salinity is measured as the percentage of pure seawater.

It is noticed that low isochlors are more sensitive to upconing than higher ones. This stems from two reasons. The first one is that the low isochlors are closer to the well, and undergo higher advective transport caused by a higher flow velocity field in the vicinity of the well. The second reason is that the denser seawater of the higher isochlors tends to prevent themselves moving upward. The latter phenomenon occurs, since the accumulative effects of variable density at the depth of the transition zone has produced a
pool with a smaller velocity beneath the salt cone. The effect of well pumping on the salt concentration distribution decreases along the $y$ direction. However, even in the vertical plane at $y = 500m$, this effect is also noticeable, especially at the top portion of the aquifer. Along the $x$ direction, the pumping well has a larger influence on the salt concentration distribution in the downstream direction than in the upstream one. This is due to the advective transport with the regional flow. As a result, the present upconing is not radially symmetric. As indicated by the position and shape of the advancing contours, the well is contaminated.

In Case C, the well is shut off at 8.27 years, when the salinity of the water pumped from the well is equal to 2%. Then, the upconed saltwater mound undergoes decay, as shown in Fig. 6. After a short time after the well’s shutoff, say 6 years, the salinity in the water in the vicinity of the well dramatically decreases, and the saltwater cone upconed to the well disappears, as shown in Fig. 6c. However, it takes a much longer time for the entire system to completely recover to the initial condition without pumping.

![Fig. 6. The decay of the saltwater cone upconing to the well and the entire system after the well is shutoff at 8.27 years in Case C.](image)

As illustrated in Fig. 7a, after a sharp decrease caused by the well pumpage, the phreatic surface remains unchanged in the upstream of the well, and undergoes a very small change in the downstream direction due to the variable density effect. However, after the pumping stops, the phreatic surface gradually approaches its initial position in both the upstream and downstream direction from the well. In this problem, we have two level iterations. The inner one is for the nonlinear nature caused by the phreatic surface. The outer one is for the coupled nonlinear density-dependent flow and transport. In each time step, two inner iterations are needed in the first iteration of the latter, then the phreatic surface does not need to be updated for the remaining iterations of the outer level, since the subsequent changes in the phreatic surface are very small.
Fig. 7b shows the temporal variation in the reference head along the bottom boundary in the $y=0m$ plane. The head globally decreases after well pumping, resulting in a larger seawater influx from the seaward side boundary. Then, the reference head remains unchanged on the side of freshwater flux ($x \leq 200m$). However, within the seawater wedge, the reference head continuously increases, as a result of interface upconing and the cumulative effect of variable density in the increasing depth of transition zone. This process increases the length of the landward flow cell and produces further seawater intrusion. The further intrusion does not stop immediately after pumping stops at 8.27 years, as illustrated in Fig. 6b and c. The well shutoff globally increases the head field. However, the competition between freshwater and seawater in the local head “valley” does not change the flow direction, since the former is dominated by the salt concentration distribution. Seawater intrusion reaches its maximum at 14 years. After that, the seawater front contracts, and the length of seawater intrusion decreases. However, this recovery process is very slow.

![Graph 1](image1)

![Graph 2](image2)

**Fig. 7.** Temporal variations in (a) the phreatic surface, and (b) the reference head at the bottom boundary in the plane of $y=0m$.

**CONCLUSION**

“Coastal collectors” are often used for pumping fresh water from the zone overlying the transition zone between fresh water and saline water close to the coast in a coastal aquifer. Local upconing of salinity contours occurs. The salinity of the pumped water depends on the rate of pumping, on the fresh water flow regime, and on the location of the well’s screen. The code *FEASi-SWIT* has been developed to simulate coupled density-dependent flow and salt transport. This code was applied to the combination of regional seawater intrusion and local upconing. In this code, the flow equation is solved by the Galerkin finite element method (FEM). The advective-dispersive salt transport equation is solved in the Eulerian-Lagrangian framework: (1) the Lagrangian concentration is derived by the adaptive pathline-based particle tracking algorithm, and (2) the remaining dispersion equation is then solved by the Galerkin FEM.
This code does not suffer from the instability constraint of Peclet number in an advection-dominated problem.

FEAsi-SWIT was applied to pumping from the transition zone in a phreatic coastal aquifer. Simulation results show that the well can pump essentially fresh water when the pumping rate is relatively small. For larger pumping rates, the salinity of the pumped water will reach the limiting value of 2% after a certain period of pumping. In this case, the intermittent pumping technique may be used for groundwater utilization.

References


**Keywords**: seawater intrusion, numerical modeling, finite element method, particle tracking

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