WELL-INFILTRATION IN FRESH-WATER POCKETS IN SANDY RIDGES IN ZEELAND
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Abstract

Because of the shortage of groundwater of good quality in summer, damage occurs to fruit trees on the peninsula of Zuid-Beveland in the southwestern part of The Netherlands. In order to investigate the possibility of enlarging the storage in existing fresh-water pockets, the Committee on Watermanagement and Desalinization performed a well-infiltration test. The well was positioned in the transition zone between fresh and saline groundwater and the infiltration test gave good results. With a non-steady numerical model, based on the theory of vortices, the movements of the interface between fresh and saline groundwater was simulated for a confined aquifer. Anisotropy was also introduced in the model. The transition zone was modelled by three layers of different densities. Calculations with the model showed the permeability and the density differences to be quite important. The upconing due to extraction from the same well as calculated did not agree very well with the results of field observations. The ratio between vertical and horizontal permeability seems to increase in periods of rest and even more in periods of extraction. This phenomenon might be caused by swelling of clay particles, local inhomogeneities and the influence of stratification. Another cause of the difference between observations and calculations might have been a non-uniform distribution of the in- and outflow over
the length of the filter. Longterm calculations showed that good results can be obtained by shallow extraction wells, but also showed the danger of depletion of the present storage by uncontrolled extraction and of quality deterioration by dispersion in the transition zone.

1 Introduction

In the province of Zeeland, The Netherlands, the yearly loss by drought of the cultivation of fruit amounts to 22 million Dutch florins. At present the farmers on the peninsula of Zuid-Beveland provide the fruit plants in summer with groundwater from present natural fresh-water pockets. In 1983 respectively 1984 the recovered quantity was 45600 respectively 31600 m$^3$. Because of the dynamic equilibrium of such small rainfed fresh-water pockets special care has to be taken to prevent depletion (Mann 1985). The Committee on Watermanagement and Desalinization (CWO) in Goes started a well-infiltration test at Kapelle (Figure 1) to investigate the possibility of creating an extra subsurface storage with collected surface water of good quality, which is only available in winter.

2 Geology and hydrogeology

2.1 Sandy ridges

The ridges are remnants of former tidal creeks and originate from the holocene transgression in the Subatlanticum, 2000 B.C. The peat layer on top of the pleistocene sands, which developed after the Weichselien glacial period, was partly eroded during this holocene transgression and creek systems were created. Behind the levels of the creeks, clay sediments settled on top of the peat formation. In later regression periods the creeks were filled up with coarse material. During the last ages the peat and clay layers in the backswamps were affected more by subsidence than the creek sediments, and the creeks nowadays appear as ridges in the countryside.
The sandy creek sediments are connected with the underlying pleistocene sands. Because of the very likely absence of semi-permeable layers the base of the aquifer is formed by marine clay from the Rupelien at about 100 m below surface.

The permeability of the aquifer material is estimated to be 12 m/d.

Figure 1. Infiltration site
2.2. Fresh-water pockets

At the ridges the clay and peat layer is absent. Percolation of rainwater slowly formed fresh-water pockets in the sandy saline soil. The recharge is about 250 mm per year. The depth of the fresh-water pocket is indicated by different shading (Figure 1). The greatest thickness is 25 to 30 m at Kapelle.

The presence of fresh groundwater and the mild climate are ideal factors for growing fruit and vegetables.

3 The well-infiltration test

The CWO constructed three well-sites. The infiltration-well IP1 was used in this study for a model-test. A system of ditches delivered the recharge for infiltration at IP1. In this way from January 1984 until May 1985 17300 m$^3$ were infiltrated. A total amount of 9300 m$^3$ was recovered in August and September 1985 before the salinity reached the ultimate value of 500 mg Cl$^-$/l.

With filters and a special geo-electrical measuring-technique the movement of the brackish transition zone was observed.

4 Modelling

4.1 The selected model

From the great numbers of computer models available a non-steady model was selected, able to calculate the movement of the interface or the transition zone between fresh and saline groundwater. To simulate the well-infiltration the principle of axial-symmetry has to be applied. A suitable model was available at the Waterworks Research Centre of KIWA in Nieuwegein. The model BUBBLE satisfies the demands mentioned above, and because of the good documentation by Peters (1983) it was ready for use. The numerical calculations are based upon the vortex theory as elaborated by de Josselin de Jong (1977) and Haitjema (1980).
4.2 Vortex theory

If density differences occur in an aquifer, fluid flow can be modelled with vortex theory. A discrete vortex is the total rotation of a partial area concentrated in the centre of that area (Figure 2).

Figure 2. A vortex on an interface

The principles of that theory are:

a) sharp interfaces between fluids of constant density;
b) equal fluid pressures at either side of an interface;
c) identical viscosity value of fluids with different density;
d) no mixing, dispersion and diffusion;
e) homogeneous aquifer;
f) incompressibility of fluid and soil.

De Josselin de Jong (1977) derived the basic equations from the Darcy formula for the relation between specific flow $q$ and pressure $p$ and the principle of continuity in terms of the flow function $\psi$.

Darcy:
\[
q_x = -\frac{\kappa}{\mu} \frac{\partial p}{\partial x}
\]
\[
q_z = -\frac{\kappa}{\mu} \left( \frac{\partial p}{\partial z} + \Psi \right)
\]

Continuity:
\[
\frac{\partial q_x}{\partial x} + \frac{\partial q_z}{\partial z} = 0
\]

Flow:
\[
q_x = -\frac{\partial \Psi}{\partial z}
\]
\[
q_z = -\frac{\partial \Psi}{\partial x}
\]
Together this gives:  \[ \nabla^2 \psi = -\frac{\kappa}{\mu} \frac{\partial \psi}{\partial x} = \lambda \]

\[ \kappa = \text{intrinsic permeability} \]
\[ \mu = \text{dynamic viscosity} \]
\[ \gamma = \text{product of density } \rho \text{ and gravity acceleration } g \]
\[ \lambda = \text{vorticity (rotation per unit area)} \]

The derived Poisson-equation reduces to the formula of Laplace in case of homogeneity and constant density.

With a tilted interface between two fluids of differing density the vortex becomes

\[ d\Omega = \lambda d\mathbf{A} = -\frac{\kappa}{\mu} (\mathbf{v}_2 - \mathbf{v}_1) \cdot \sin \alpha \cdot b \cdot ds = \omega \cdot ds \]

with  
\[ \Omega = \text{vortex} \]
\[ b = \text{width of an interface element} \]
\[ s = \text{distance along interface} \]
\[ \alpha = \text{angle of inclination of interface} \]
\[ \omega = \text{vorticity (rotation per unit interface length)} \]

The thickness of an element of the interface disappears in the deviations; so the discontinuity at the sharp interface gives no problems. The tilted element of the interface contributes to the flow in a certain point \( P \) of the aquifer according to the equation of Lamb (Figure 3).

Figure 3. Equation of Lamb
\[ |dq^p| = \frac{\omega \cdot ds}{4\pi R^2} \sin \theta \]

with

- \( R \) = length of the vector from interface vortex element to the point in the aquifer considered;
- \( \theta \) = angle between rotation axis and vector

Elaborating these equations, the influence of all the elements of the interface must be taken into account. In case of axial-symmetry this can be presented in the form of vortex-rings (Haitjema 1980).

A complete elliptical integral results that can be solved numerically.

### 4.3 Well-infiltration

The influence of the well in the aquifer can be calculated assuming the well, with total filterlength 2L and infiltration quantity \( Q \), is built up with point wells with height \( dz_0 \) (Figure 4).

\[ dq = \frac{Q}{2L} \cdot dz_0 \]

For the specific flow in the aquifer at a distance \( r \) from a point well can be derived:

\[ Q = 4\pi r^2 (-k \frac{\partial \phi}{\partial r}) \]

![Figure 4. Well infiltration](image-url)
Integrating over the height of the filter gives for the actual flow \( n \) (porosity):

\[
\begin{align*}
V_{r,act} &= \frac{Q}{8\pi L n} \frac{1}{r} \left[ \frac{L-Z}{\sqrt{(L-Z)^2 + r^2}} + \frac{L+Z}{\sqrt{(L+Z)^2 + r^2}} \right] \\
V_{z,act} &= \frac{Q}{8\pi L n} \left[ \frac{1}{\sqrt{(L-Z)^2 + r^2}} - \frac{1}{\sqrt{(L+Z)^2 + r^2}} \right]
\end{align*}
\]

In this way the exact location of the infiltration well from a test-site can be simulated with the model.

4.3 The BUBBLE-model

A computer program was written by Peters in FORTRAN representing, in a main program and a series of subroutines, the above mentioned formulae by superposition of density and infiltration effects. The model calculates the movement of one or more (if desired) interfaces in a semi-infinite homogeneous and isotropic aquifer by introducing an image well and interfaces at the other side of the impervious top layer.

An updating procedure prevents clustering or sweeping away of interface nodal points. After each timestep in the calculation process the new position of a nodal point is corrected to a place just above or below the former position.

4.4 Anisotropy

By the nature of sedimentation processes the assumption of isotropy is not reasonable. Therefore the intrinsic permeability will have different values in horizontal and vertical direction. With the assistance of van Duijn, Mathematics Department of Delft University of Technology, a transformation factor \( \alpha \) as proposed by de Josselin de Jong (1977) was used in the model to simulate an anisotropic aquifer.
By transforming the anisotropic prototype \((x, z)\) an anisotropic equivalent \((r, \theta)\) domain is created with as isotropic intrinsic permeability the horizontal component of the anisotropic intrinsic permeability \(k_H\).

\[
\begin{align*}
q_x &= -\frac{k_H}{\mu} \frac{\partial p}{\partial x} \\
q_z &= -\frac{k_V}{\mu} \left[ \frac{\partial p}{\partial z} + \nu \right] \\
\theta &= \sqrt{\frac{k_H}{k_V}} z \\
\phi &= \chi
\end{align*}
\]

The vertical dimensions of aquifer, well and interfaces must be multiplied by the transformation factor. One can think of a cilindric domain stretched out in vertical direction. Also the total infiltration quantity must be transformed with the same factor.

Some attention must be paid to the aquifer and fluid properties. From the transformation derivations it follows that the density has to be divided by the transformation factor

\[
\tilde{\rho}(r, z) = \rho(x, \sqrt{\frac{k_H}{k_V}} z) = \rho(x, z)
\]

\[
\begin{align*}
\tilde{q}_r(r, z) &= q_x(x, z) \\
\tilde{q}_\theta(r, z) &= \alpha q_z(x, z) \\
\text{div} \tilde{\varphi} &= 0 \\
\tilde{q}_r &= -\frac{k_H}{\mu} \frac{\partial \tilde{p}}{\partial r} \\
\tilde{q}_z &= -\frac{k_H}{\mu} \left[ \frac{\partial \tilde{p}}{\partial z} + \sqrt{\frac{k_V}{k_H}} \nu \right] = -\frac{k_H}{\mu} \left[ \frac{\partial \tilde{p}}{\partial z} + \nu \right]
\end{align*}
\]

The permeability \(k\) has to be transformed proportional to the density according to the well-known relation

\[
k = \frac{\rho \varrho g}{\mu}
\]

4.5 Stability criterium

To achieve a stable calculation process Peters formulated a criterium for the relation between the timestep \(\Delta t\) and the distance \(d\) between interface nodal points.
\[ \Delta t < \frac{2\Delta \rho}{k \Delta \rho} \]

with
\[ \rho = \text{density} \]
\[ \Delta \rho = \text{density difference between fluids} \]
The timestep can be chosen longer with greater \( d \), the choice depends on the desired precision of modelling the interfaces.

4.6 Influence of permeability and density values

In Figure 4 and Table 1 the results of some calculations with variation of geohydrological constants are represented for a well-infiltration of 120 m\(^3\)/d during 152 days and a recovery rate of 180 m\(^3\)/d.

One can conclude that the influence of anisotropy is important (cases A, B, C). Also the effects of deviating estimations of constants may be considerable (cases A, D, E).

Table 1.

<table>
<thead>
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<th></th>
<th>( k_H )</th>
<th>( k_V )</th>
<th>( \alpha )</th>
<th>( \rho_s )</th>
<th>( \rho_f )</th>
<th>( z )</th>
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<td>3</td>
<td>2</td>
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<td>28</td>
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<tr>
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</table>

\( z \) = interface displacement at the well after infiltration period
\( t \) = length of recovery period until interface reaches the well
\( R \) = rate of infiltrated and recovered quantity

5 Application of the model

5.1 The infiltration-test at Kapelle

In reality there exists rather a transition zone than a sharp interface between fluids with different salinity and density. For this reason the transition zone, known from measurements, at Kapelle was modelled by introducing three interfaces. Figure 5 shows this scheme, and most of
the required input data. Other input data concern the number of nodal points, timesteps and plot information.

Working with a semi-infinite confined aquifer model reasonable approximations are found of the displacements of the interface in a phreatic aquifer, varying the estimation of vertical permeability component and comparing calculations and measurements. In the model, calculations start with an initial equilibrium of the interfaces. Natural flow is neglected. The effect of the deep aquifer base is also assumed to be negligible.
5.2 Discussion of results

The calculations were compared with the salinity profiles measured at successive timesteps in the way of Figure 6. According to the comparison process it seemed that in the time between infiltration periods and in a period of recovery in the executed well-test the extra storage near the well did not decrease as quickly as in the computations. Several reasons may be stated as explanation:

a) the scale of the test. The infiltrated quantity is rather small, so the effects of inhomogeneities can be important. These may be:
b) the underestimated effect of stratification;
c) reduction of the vertical permeability component by clayswelling, due to the possible occurrence of the clay mineral montmorillonite in thin Duinkerke-II silty sediment layers;
d) deviation of the assumption of uniform flow towards the well in a recovery period;
e) effects of dispersion;
f) effects of temperature difference between infiltrated water and original groundwater on the viscosity.

In searching the best estimation of the vertical permeability it seems that to the end of the test, anisotropy becomes more pronounced, with a value for the vertical permeability varying from 9 to 3 m/d. Figure 7 shows the results of the computation. It was not possible to detect the movement of the lowest schematized interface, because of the positions of the permanent electrode systems. So changes in the width of the transition zone were difficult to detect, especially as during recovery upconing can be expected. Nevertheless the model gives good insight in the movement of interfaces and can serve to make safe predictions of future situations.

From some long term runs about the well system IPl the conclusion was drawn that it is unfavourable to use wells with the lower part of the transition zone. The maximum yield of groundwater of good quality is—even after 4 years—even 30 percent of the yearly infiltrated quantity. The use of long infiltration wells and shallow recovery wells gave good prospects (Figure 8). Although the model is well suited to cope with a non-uniform in- and outflow of the well, due to a lack of data, this could not be taken into account. This aspect needs to be investigated further.
Figure 6. Principle of the calibration method
Figure 7. Calibration
R = distance from well centre
z = depth of interface below NAP
\( t = \) time in days

Figure 8. Infiltration/recovery scheme

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