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A SENSITIVITY ANALYSIS OF THE ADAPTED GROUNDWATER MODEL MOC

G.H.P. Oude Essink
Delft University of Technology, Faculty of Civil Engineering,
Sanitary Engineering & Water Management Group,
2628 CN Delft,
THE NETHERLANDS.

SUMMARY

The two-dimensional model *MOC* of Konikow & Bredehoeft, which is originally designed for horizontal groundwater flow, is subject and means of research in order to analyse salinisation of the subsoil. The present research is carried out in the framework of a project focused on impacts of sea level changes. MOC is adapted for *density differences* with documentation of Lebbe and Van der Eem. As a result, density dependent groundwater flow with solute transport and hydrodynamic dispersion can be simulated.

In this paper, various model parameters are examined to analyse their influence on the solution. Their relation to causes of numerical instability in the velocity field is investigated. Moreover, a sensitivity analysis of some geohydrological parameters is executed for a cross-section perpendicular to the Dutch coast.

Numerical instabilities due to density flow arise from the numerical approximation and may occur at places where the transition zone between fresh and saline groundwater is small. Numerical instability can be diminished by adapting model parameters, e.g., by enlarging the number of particles per grid cell, by decreasing the duration of the flow time step or by smoothening the initial density differences at transition zones. However, enlarging the number of grid cells does not necessarily diminish numerical instability. Furthermore, geohydrological parameters, such as dispersivities and the presence of a loam layer, highly determine the salinisation process in this specific groundwater flow system.

1. INTRODUCTION

The two-dimensional groundwater flow model MOC of Konikow & Bredehoeft of the US Geological Survey is used to execute a sensitivity analysis of this groundwater flow system. The original model is adapted for *density differences*. As a result, *vertical* two-dimensional groundwater flow with solute transport and hydrodynamic dispersion can be simulated.

In this paper a cross-section perpendicular to the Dutch coast is simulated, pursuing a study of Kooiman (1988) [1]. He simulated the development of the freshwater lens in this area from the end of the year 1853 up till the end of the year 1987. Furthermore, he estimated effects of human activities, such as the reclamation of the Haarlemmermeer, groundwater abstraction for public water supply, artificial recharge and deep-well infiltration.

The objectives of the present study are to examine the various model parameters used in MOC, in order to estimate their influence on the solution of the problem and the relation of these model parameters to causes of numerical instability in vertical velocity. Second, the

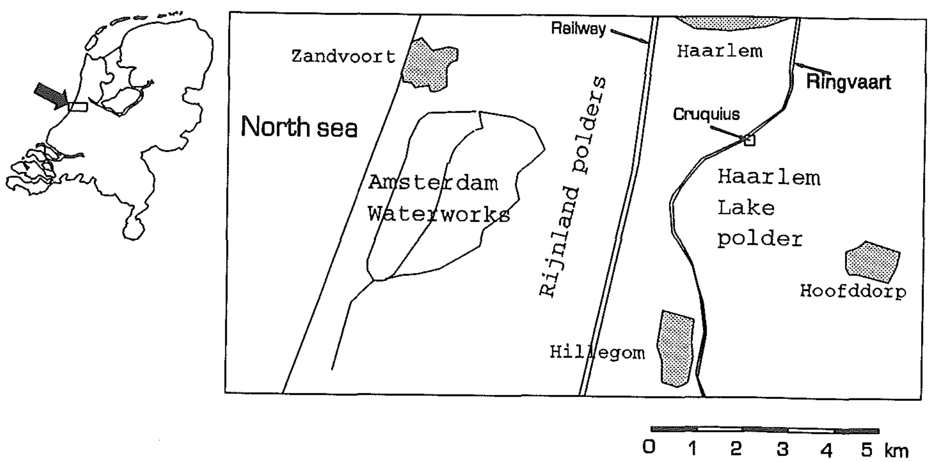

Figure 1: *The map with the sand-dune area of the Amsterdam Waterworks, the Rijnland polders and the Haarlemmermeer polder.*

influence of some geohydrological parameters are investigated, as there are still uncertainties in the actual values.

In section 2, the present situation of the cross-section is described. In section 3 values of some relevant parameters of MOC are summarized and the geohydrologic system is schematized. The causes of numerical instability are discussed in section 4 by means of examination of various model parameters. A geohydrological parameter analysis is executed in section 5, where specific storativity, dispersivities and the influence of a loam layer are argued. Finally, the conclusions are presented.

2. PRESENT SITUATION

The sand-dune area of the Amsterdam Waterworks is situated along the North Sea coast (fig. 1), which has a surface of roughly 36 km^2: an average length of 9 km parallel to the coastline and an average width of 4 km perpendicular to the coastline. In the sand-dune area the phreatic groundwater level fluctuates around a constant level. Here, the maximum difference in phreatic groundwater level is small because of the presence of many infiltration gullies, drains and canals. The Rijnland polders are situated inland where horticulture is practised, especially bulb farms. The phreatic groundwater level in this area is between -0.6 till -1.0 m *N.A.P.*[1]. The low-lying Haarlemmermeer polder, which is separated from these polders by the canal 'De Ringvaart', has a controlled phreatic groundwater level varying between -4.5 and -6.0 m *N.A.P.*

A short description of the evolution to the present situation in terms of solute distribution is as follows [1], (Schuurmans) [2]. During many centuries, natural groundwater recharge

[1] *N.A.P.* ('Normaal Amsterdams Peil') is the reference level in The Netherlands. *N.A.P.* roughly equals Mean Sea Level.

has built up a freshwater lens in the geohydrologic system along the coast. Before 1854, the situation of the freshwater lens was supposed to be stationary (Venhuizen) [3]. In 1854, the Haarlemmermeer was reclaimed, which resulted in a serious salt water intrusion from the sea. Meanwhile, the Amsterdam Waterworks had started the abstraction of raw water from the sand-dune area. During the 49 years that followed, groundwater was withdrawn from the phreatic aquifer only. From 1902 up till 1956, shortage of drinking water made it necessary to abstract groundwater also from the deeper semi-confined aquifer. The abstraction of groundwater by deep-wells increased from a few million m^3 per year to nearly 18 million m^3 per year in the period 1951-1956. The once 'sharp' salt-fresh interface underneath the sand-dunes transformed into a brackish transition zone. To compensate these negative consequences, the Amsterdam Waterworks started in 1957 with the infiltration of riverwater (Rhine water) on the surface, while the deep-well abstractions were reduced to 6 million m^3 per year. Until 1974, the situation of the freshwater lens did not change significantly. From 1974 on, the deep-well abstraction is diminished to an average of 1.75 million m^3 per year. During the past 19 years, effects of infiltration of fresh water in the aquifers (*deep-well infiltration*) has been studied. With this recharge, the Amsterdam Waterworks can increase the volume of the freshwater lens and counteract the threatening salt water intrusion underneath sand-dunes and polders.

3. SCHEMATIC OUTLINE

In this section, first a short description of MOC is given, followed by a summary of model and geohydrological parameters used in the reference case. Furthermore, the geometry of the cross-section is given. The data of the schematization of the cross-section have been borrowed from the following sources: Kooiman [1], Schuurmans [2], Venhuizen [3], Stuyfzand [4], ICW [5] and TNO [6].

3.1 Short description of MOC

The solute transport model MOC ('Method of Characteristics') from the US Geological Survey is developed by Konikow & Bredehoeft in 1978 [7]. Originally, MOC is a *horizontal* two-dimensional groundwater flow model. However, *vertical* cross-sections have to be simulated here, since the density distribution in the groundwater flow system is non-uniform. The groundwater flow equation in MOC (version 3.0 of 1989, [8]) is adapted for *density differences*: see Lebbe [9], [10], Van der Eem [11] and Oude Essink [12].

MOC is suited to simulate non-conservative solute transport in saturated groundwater systems. The adapted model can compute changes in spatial density distribution over time caused by convective transport, hydrodynamic dispersion, dilution from recharge and physical processes, such as absorption and decay. However, in this paper, physical processes are left out of consideration. Since in MOC freshwater heads have to be applied, the existing piezometric levels (fresh, brackish or saline) are converted into freshwater heads.

The quality of groundwater is determined by dissolved solutes. These dissolved solutes are divided into negative (e.g., Cl^-, SO_4^{2-}) and positive ions (e.g., Na^+, Mg^{2+}). Since the chloride content in groundwater is the most important negative ion and is investigated intensively, the interest in this paper is focused on the distribution of these chloride ions. Therefore, when here in fact only changes in chloride distribution are simulated, the distribution of all dissolved solutes and salt water intrusion are meant. The classification of *fresh*, *brackish* and

Type of groundwater	Chloride concentration [$mg\ Cl^-/l$]
fresh	$Cl^- <= 300$
brackish	$300 < Cl^- < 10000$
saline	$Cl^- >= 10000$

Table 1: *The classification of fresh, brackish or saline groundwater depending on the chloride concentration, after Stuyfzand.*

saline groundwater is according to Stuyfzand [4], see table 1. In MOC, density flow is simulated, and thus the density distribution in the entire geohydrologic system must be known. Density is a function of pressure, temperature and the concentration of dissolved solutes. However, under the given circumstances of the simulated geohydrologic system, the influence of both pressure and temperature can be neglected. Therefore, in this paper the density of groundwater is only related to the concentration of dissolved solutes in the groundwater. The conversion from chloride concentration to density, which is used in the adapted MOC, is as follows (equation 1):

$$\rho_i = \rho_0 * (1 + \frac{\rho_{ref} - \rho_0}{\rho_0} * \frac{C_i}{C_{ref}}) \tag{1}$$

where ρ_i is the density of groundwater at grid cell i [ML^{-3}]; ρ_0 is the standard reference density, usually the density of fresh groundwater ($\rho_0 = 1000\ kg/m^3$); ρ_{ref} equals the density of saline groundwater ($\rho_{ref} = 1025\ kg/m^3$); $\frac{\rho_{ref}-\rho_0}{\rho_0}$ is the relative density difference [$-$] ($\frac{\rho_{ref}-\rho_0}{\rho_0}$=0.025); C_i is the chloride concentration in grid cell i [$mg\ Cl^-/l$]; and C_{ref} is the reference chloride concentration for which saline groundwater equals a density of 1025 kg/m^3 ($C_{ref} = 18630\ mg\ Cl^-/l$).

MOC employs the method of characteristics to solve the solute transport equation. It applies a *particle tracking* procedure to represent convective transport and a two-step explicit procedure to solve the finite difference equation, that describes the effects of hydrodynamic dispersion and fluid sources and sinks. *Particle tracking* means that a number of particles is placed in each grid cell and traced during the groundwater flow. Each particle represents a certain chloride concentration. After the particles are moved according to the groundwater flow, particles can have entered other grid cells. At the end of a certain time increment, the averaged chloride concentration of every grid cell is computed again with the available particles in that grid cell. Subsequently, the chloride concentration of each particle is adjusted to the averaged chloride concentration of the grid cell. The explicit *particle tracking* procedure is subject to stability criteria, but the program automatically determines and implements the time step limitation necessary to satisfy the stability criteria. This time increment, that is used in the solute transport equation, is called the *chemical time step*[2].

3.2 Parameters in the reference case

(a) Model parameters

Total number of grid cells is 4500: 250 in horizontal direction and 18 in vertical direction. A grid cell is 50 m long and 10 m high, thus the total simulated cross-section is 12500 m long and 180 m high. Each grid cell contains initially five particles. The convergence criterion

[2]A *chemical time step* is the time increment in which a particle is moved over a distance proportional to the length of that time increment and the velocity at the position of that particle.

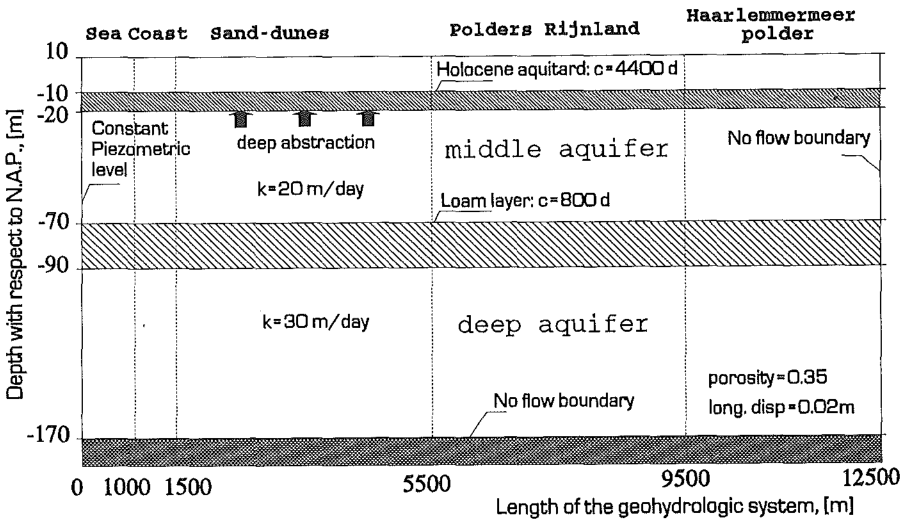

Figure 2: *Geohydrological parameters in the cross-section of the Amsterdam Waterworks and the polders lying inland.*

TOL for the iterative calculation of the groundwater flow equation is 10^{-5} $feet^3$. The maximum distance CELDIS across one grid cell, in which a particle is allowed to move during one *chemical time step*, is set to 0.5 (as a fraction of the grid cell). After each *flow time step* of one year, the groundwater flow equation is calculated again. During this flow time step, the velocity field remains constant. In the reference case, several *chemical time steps* are necessary to simulate the solute transport during the flow time step of the groundwater flow. The saturated thickness of the aquifer represents in the adapted model a parameter that determines the thickness of the aquifer *perpendicular* to the cross-section. This parameter is set to the unity length of one foot.

(b) Geohydrological parameters

In the cross-section given by Kooiman [1], the geohydrological parameters (fig. 2) and the phreatic groundwater level are averaged *perpendicular* to the cross-section. Fig. 3 shows the initial chloride distribution at the year 1854, as proposed by Kooiman [1].

The determination of the longitudinal dispersivity α_L gives some difficulties. The mechanical mixing process of dispersion (hydrodynamic dispersion) not only depends on the velocity field distribution, but also on the geometry of the geohydrologic systems with its good permeable aquifers and less permeable aquitards. According to literature studies such as Bertsch [13], laboratory dispersion experiments report measurements of the longitudinal dispersivity α_L in the range of 10^{-4} to 10^{-2} m. By contrast, field experiments in heterogeneous sand and gravel aquifers (especially in the USA) are calibrated with values of α_L in the range of 1 to 100 m. However, in Dutch and Belgian groundwater flow systems, the best estimations appear to be rather small. Considering a few case-studies (such as [14], [15] and [16]), the longitudinal dispersivity α_L is initially set here to a small value: 0.02 m. A limited sensitivity analysis on this parameter is executed in this paper. The ratio transversal

[3] MOC uses the Anglo-Saxon system.

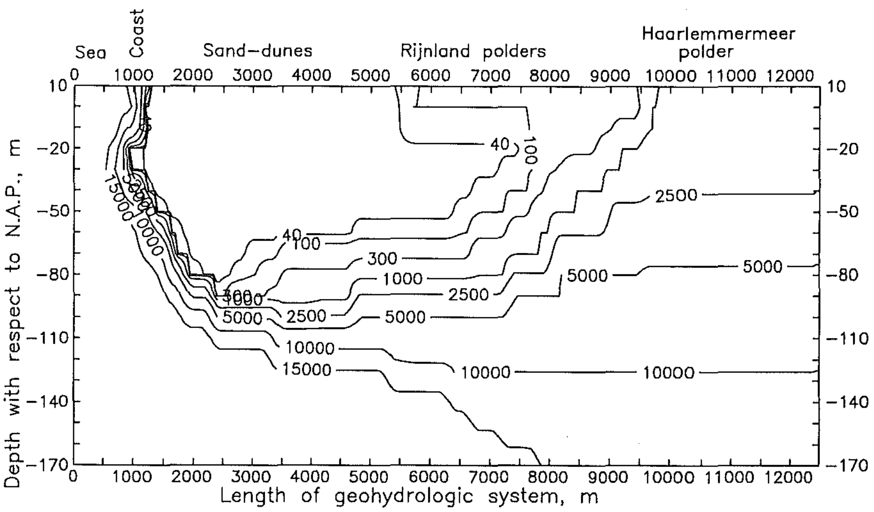

Figure 3: *The initial chloride distribution at the year 1854 (Kooiman, 1988).*

to longitudinal dispersivity is set to 0.1 [–] and the anisotropy factor (the ratio vertical and horizontal hydraulic conductivity) is also 0.1 [–]. The effective porosity is 0.35 [–] and the specific storativity S_s is equal to zero.

(c) Boundary conditions

At the horizontal upper boundary, the phreatic groundwater level in the Haarlemmermeer polder and the Rijnland polders are kept constant at respectively -5.50 m *N.A.P.* and -0.60 m *N.A.P.* The natural groundwater recharge into the phreatic aquifer at the sand-dunes is 360 mm per year. A *constant piezometric level* boundary is introduced at the seaside. As the piezometric level in the upper grid cell and the density distribution in vertical direction are known, all grid cells underneath the upper cell in the same column are corrected for density differences. For every grid cell at an increasing depth with respect to *N.A.P.*, the freshwater head at the centre of the grid cell i is as follows (equation 2), taking into account that the situation is supposed to be *hydrostatic*:

$$\phi_i = \phi_{i-1} + \frac{\rho_{av,i} - \rho_0}{\rho_0} * \Delta h \qquad (2)$$

where ϕ_i is the freshwater head in grid cell i $[L]$; $\rho_{av,i}$ is the averaged density of groundwater between the two neighbouring grid cells i and $i-1$ $[ML^{-3}]$; and Δh is the height of a grid cell (10 m).

(d) Calibration

The calibration of the simulations for this cross-section is done -not shown in this paper- for four different parameters: chloride distribution at certain points of time in the past; piezometric levels at some depths relative to *N.A.P.*, seepage through the Holocene aquitard in the Haarlemmermeer polder; and natural groundwater recharge in the sand-dunes.

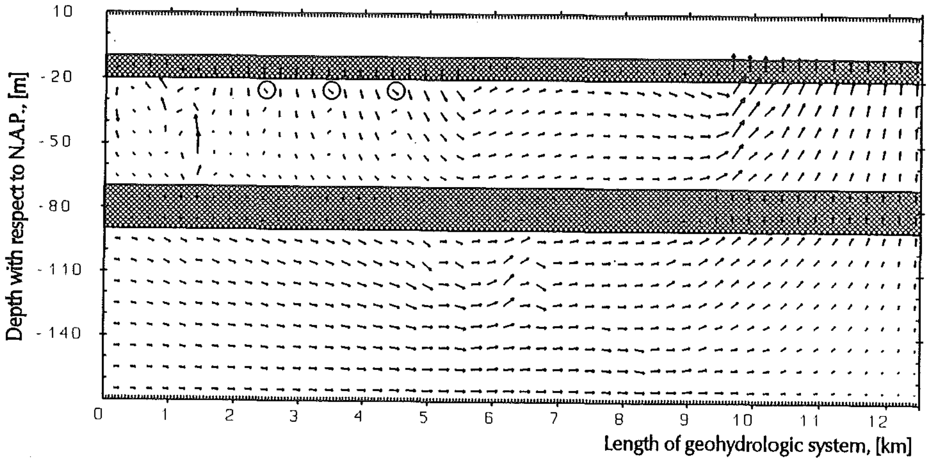

Figure 4: *The velocity field of the geohydrologic system in 1988. The length of the arrow represents the displacement of a groundwater particle during 10 years, if the velocity field would be constant during that time.*

4. CAUSES OF NUMERICAL INSTABILITY

In this section the causes of numerical instability in the velocity field and relevant parameters concerning this phenomenon are discussed. Because *density differences* are taken into account in the adapted MOC, the vertical velocities and thus the groundwater flow equation both depend on the density distribution over the grid cells. In MOC velocities can be given both into the centres of grid cells, as well as on the boundaries between grid cells. If numerical instability occurs, it depends among others on the geometry of the geohydrologic system whether the vertical velocities in the centres of grid cells fluctuate in direction with approximately equal amplitudes, or diverge to unacceptable great amplitudes.

For illustration, fig. 4 shows the velocities through the geohydrologic system in 1988, 134 years after the reclamation of the Haarlemmermeer in 1854. The velocity in the centre of every fifth horizontal grid cell is displayed. There are three groundwater abstraction wells at the sand-dune area. As can be seen, salt water intrudes into the *deep* aquifer, while vertical velocities dominate in the resistance layers. In the sand-dunes, there is a downward groundwater flow through the Holocene aquitard (natural groundwater recharge), while in the Haarlemmermeer polder groundwater flows upward (seepage). Especially in the *middle* aquifer under the coast at the first one and a half kilometre, the velocities have large fluctuations in vertical direction. In this area, a small transition zone between fresh and saline groundwater is present (see fig. 3). The fluctuations seem to be very great because the vertical dimension is magnified with a factor thirty-five.

However, fig. 5 and fig. 6 show the same velocity field with smaller magnifications of the vertical dimension. The velocity in the centre of every horizontal grid cell is displayed. Now, the fluctuating vertical velocities appear to be small, compared to the dominant horizontal velocities of groundwater in the geohydrologic system.

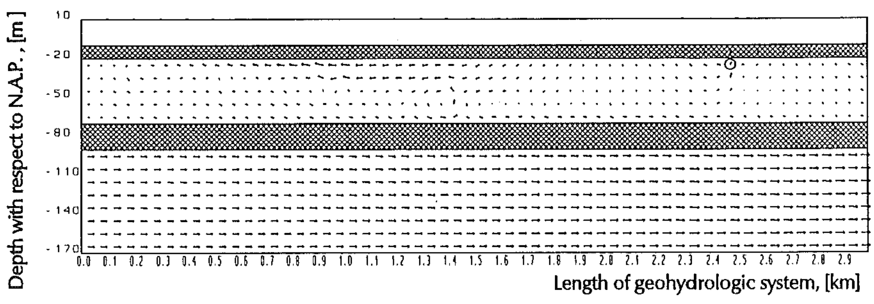

Figure 5: *The velocity field in the year 1988 of the first three kilometres with the magnification of the vertical direction equal to a factor five. The length of the arrow represents the displacement of a groundwater particle during 3 years.*

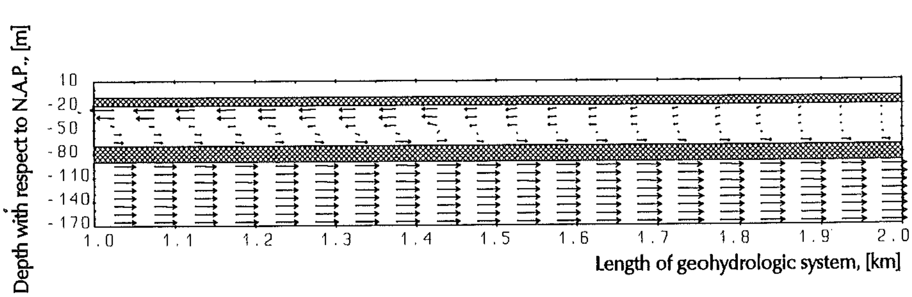

Figure 6: *The velocity field in the year 1988 between the first and second kilometre of the geohydrologic system without a magnification of the vertical direction. The length of the arrow represents the displacement of a groundwater particle during 3 years.*

The density of a certain grid cell, that contains particles with a low chloride concentration, can change abruptly if particles with a high chloride concentration enter that certain grid cell from a neighbouring grid cell, taking into account that equation 1 is used to convert chloride concentration of a grid cell into density. Abrupt differences in density between neighboring grid cells due to particle movement may cause changes in the vertical velocity. Sometimes the direction of vertical velocity may even alter, at least if the base vertical velocities are small. These differences in direction of the vertical velocity may cause numerical instabilities. As a result, numerical instability may occur in the vertical velocity field at those places, where the density in neighboring grid cells differs strongly. For example, major changes may occur at small transition zones between fresh and saline groundwater or at places of deep-well infiltration where surface water with a density, different from that of the groundwater in that area, is injected. The causes of numerical instability are closely connected with the following model parameters as well as with the hydrodynamic dispersion.

4.1 Number of particles in each grid cell

Numerical instability may occur if a particle with a high chloride concentration -and thus a high density[4]- enters a grid cell, where the already present particles have a low chloride concentration -and thus a low density-. Then, after averaging the chloride concentration -and thus the density- over the available particles at that moment, the increase in density is less sudden if more particles per grid cell are present. As a result, more particles per grid cell counteract numerical instability and make the simulation of the solute transport more accurate.

In fig. 7, the fluctuations of the chloride concentration (as a function of time) are given for cases with some different model parameters. The conversion from density into chloride concentration is also given in equation 1. Other parameters are equal to the values in the reference case. The chloride concentration as a function of time of the reference case has quite some fluctuations, while in the cases with 9 and 16 particles (the *accurate* case), the chloride concentration is more smoothened.

4.2 Convergence criterion TOL

If the groundwater flow equation is solved more precise -TOL is smaller-, the velocity field and the particles movements will be more accurate. As a result, numerical instability is counteracted, although this parameter has less influence on the accuracy of the solute transport than the number of particles per grid cell (see fig. 7).

4.3 Duration of flow time step, dt

During one flow time step the velocity field is kept constant, while particles are displaced at *chemical time steps*. If too many *chemical time steps* are simulated in one flow time step, the density distribution can be changed too much to fit with the solution of the groundwater flow equation. As a consequence, numerical instability may be created. To limit this effect, the duration of the flow time step should be decreased.

The differences in chloride concentration between the reference case and the *accurate* case are mostly smaller than 100 $mg\ Cl^-/l$, when the 134 years after the reclamation of the Haarlemmermeer polder are simulated. However, in the surroundings of the transition zone under the sand-dunes, the difference on chloride concentration can be up to a few thousand $mg\ Cl^-/l$ in only a few grid cells.

4.4 Initial density distribution

The simulation of the geohydrologic system starts with an initial density distribution. The density distribution is related to the chloride distribution according to equation 1. If the initial density distribution has vertically neighbouring grid cells with a relatively great difference in density, this may immediately create numerical instabilities in the velocity field. Therefore, the initial chloride distribution must be as smooth as possible. Of course, the simulated chloride distribution must be in accordance with the measured distribution.

4.5 Dimension of grid cells and total number of grid cells

If more grid cells represent the same geohydrologic system, these grid cells have smaller dimensions. The groundwater flow equation is also solved more accurately and the particle

[4] The conversion from chloride concentration to density is given by equation 1.

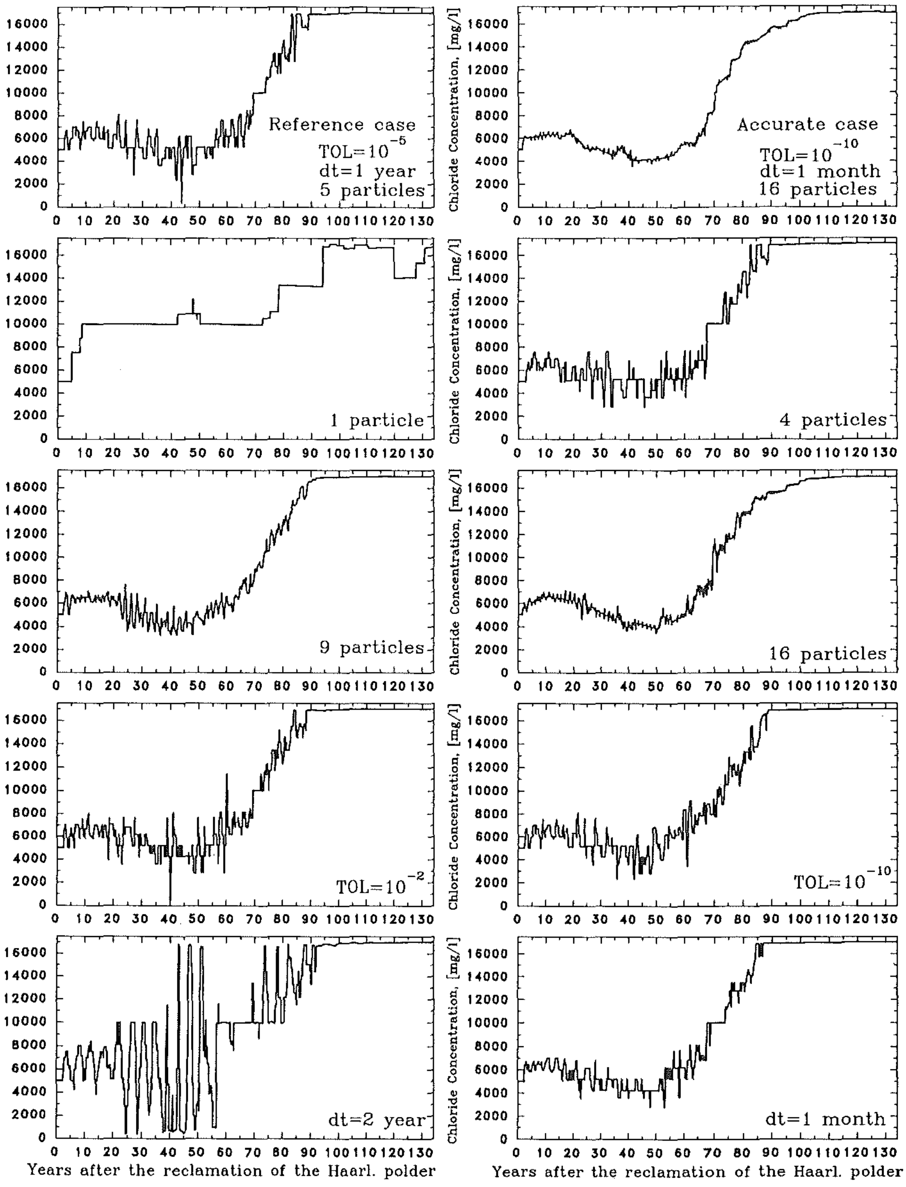

Figure 7: *The chloride concentration as a function of time in the observation well at x=3475 m and y=-105 m N.A.P. The influence of different model parameters is shown.*

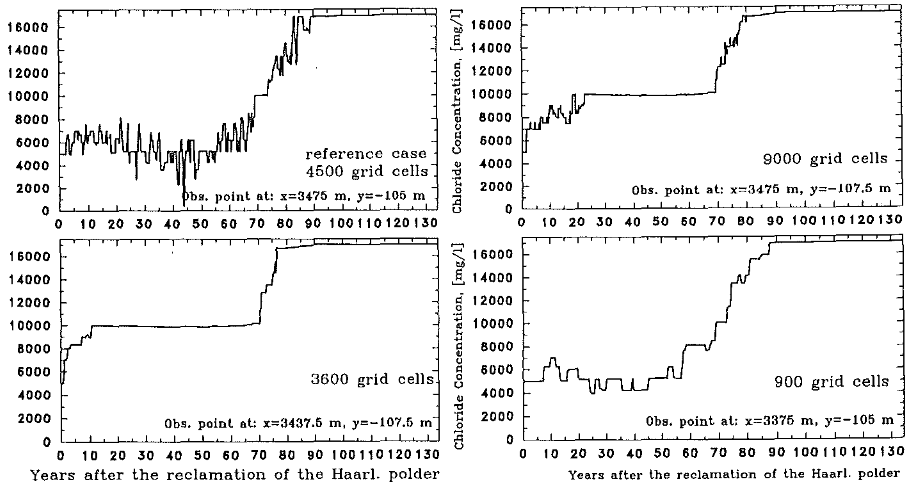

Figure 8: *The chloride concentration versus the time (in years after 1854) in an observation well. The influence of different numbers of grid cells is shown.*

movements more precisely. More grid cells means more particles, and therefore, the solute transport is also simulated more accurately. On the other hand, if the grid cells are too small, a particle in an instability zone can more easily enter another neighbouring grid cell with a completely different density. Thus, the numerical instability may be amplified. However, in this particular case the favorable characteristics of smaller -and thus more- grid cells dominate and numerical instability is counteracted.

In fig. 8 the chloride concentrations as a function of time in observation wells for four different numbers of grid cells are compared. The observation wells are located in a small transition zone. The following cases with different numbers of grid cells for the same geohydrologic system have been simulated: (a) reference case 250*18=4500 grid cells; (b) 50*18=900 grid cells; (c) 100*36=3600 grid cells; and (d) 250*36=9000 grid cells. The observation wells of the three new cases have slightly different positions compared to the reference case, because the observation well is always situated in the centre of a grid cell.

As can be seen, the reference case has the most severe fluctuations in chloride concentration. So, in this particular situation, cases with less grid cells -900 and 3600 grid cells- does not necessarily lead to a less accurate solution. If larger grid cells in horizontal direction are used in the simulation, less particles are moved to neighboring grid cells, and as a consequence, there are less abrupt changes in chloride concentration -and thus in density- after averaging over the available particles. For instance, the case with 900 grid cells -thus less grid cells in horizontal direction- resembles the reference case most, only with less severe fluctuations. Therefore, severe fluctuations in chloride concentration in those grid cells may decrease.

However, as can be seen by comparing the reference case -4500 grid cells- to the case with 9000 grid cells, more grid cells in vertical direction reduces the fluctuations. Especially in a transition zone, abrupt changes in density between neighboring grid cells are smaller, and

Number of grid cells	Nat. Rech. Sand-dunes $10^3 m^2 yr^{-1}$	Seepage Haarl. meer $10^3 m^2 yr^{-1}$	Volumes over geohydrologic system: Fresh $10^3 m^2$	%	Brackish $10^3 m^2$	%	Saline $10^3 m^2$	%
900	573.3	655.6	168.0	24.0	196.0	28.0	336.0	48.0
3600	560.0	653.3	165.1	23.6	198.4	28.3	336.5	48.1
4500	572.2	650.0	172.8	24.7	194.6	27.8	332.6	47.5
9000	562.2	650.0	165.8	23.7	199.1	28.4	335.1	47.9

Table 2: *Influence of cases with different grid cells on some geohydrological parameters.*

numerical instability is counteracted. No general recommendation can be given for the ideal number of grid cells in horizontal direction, since the occurrence of numerical instability depend on both the geometry of the geohydrologic system as well as on the location of transition zones.

In table 2, some results of the four cases with different grid cells are given. The volumes of fresh, brackish and saline groundwater are in million m^3 per metre width, calculated for the geohydrologic system in the year 1988: in horizontal direction between 0 and 12500 m, and in vertical direction between -10 and -170 m *N.A.P.* The groundwater flow at the sand-dunes, between 1500 and 5500 m and through the Holocene aquitard, represents the natural groundwater recharge, and the groundwater flow in the Haarlemmermeer polder through the Holocene aquitard represents the seepage. Altogether, the results are very similar.

4.6 Hydrodynamic dispersion

The hydrodynamic dispersion, by means of longitudinal and transversal dispersivities α_L and α_T, influences also the numerical instability. If dispersivities are small, mixing between neighbouring grid cells will be small too. Then it is possible that a great difference in chloride concentration -and thus density- can occur between neighbouring grid cells, and as a result, this may cause numerical instability.

5. GEOHYDROLOGICAL PARAMETER ANALYSIS

In this section the following geohydrological parameters, that influence the accuracy of the simulation of the geohydrologic system, are shortly discussed: the specific storativity S_s; the longitudinal dispersivity α_L; and the presence of the loam layer between -70 m and -90 m *N.A.P.*

5.1 Specific storativity S_s

The time span, in which a (dynamic) steady state situation of the piezometric level distribution in the geohydrologic system is attained, depends among others on the value of the specific storativity S_s[5]. Computations with MOC are simplified significantly when it is possible to use a specific storativity S_s equal to zero. Therefore, several simulations will show that the time scale, used in the case-study over which changes in piezometric level are accomplished in this relative extended geohydrologic system, is large enough to justify that the specific storativity S_s may be taken equal to zero.

[5] As the adapted version of MOC is used, the saturated thickness b of the aquifer *perpendicular* to the cross-section is set to the unity length of 1.0 foot. Therefore, the storage coefficient S [$-$] is equal to the specific storativity S_s [L^{-1}]: $S = S_s * b$, where b equals the thickness of the aquifer.

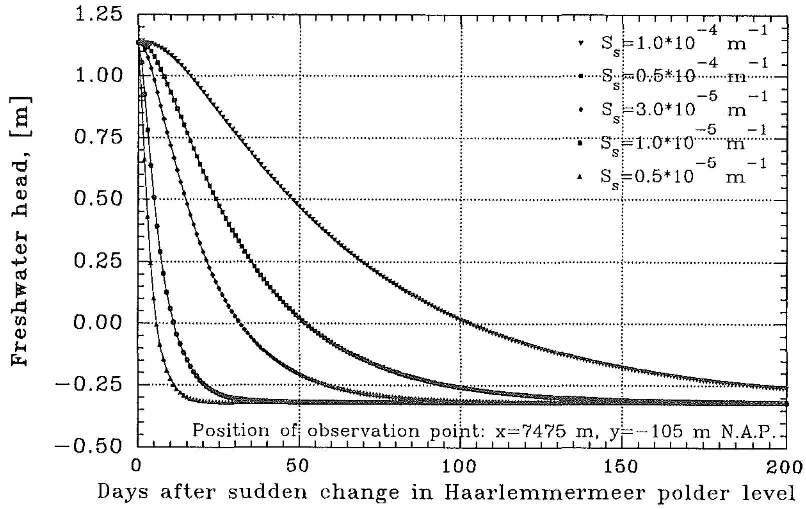

Figure 9: *The influence of the specific storativity S_s on the freshwater head as a function of time in a observation well with the position x=7475 m and y=-105 m N.A.P., in case the phreatic groundwater level in the Haarlemmermeer polder is changed suddenly. The flow time step is one year.*

The values of the specific storativity S_s that are obtained from pumping tests in the area of the Amsterdam Waterworks, are in the range of $1 * 10^{-5} m^{-1}$ to $3 * 10^{-5} m^{-1}$ (Uffink) [17]. Furthermore, an empirically determined formula, developed by Van der Gun [18] who used results of pumping tests of almost sixty cases in Dutch sandy aquifers, also generates values of the specific storativity S_s of the geohydrologic system in the same range.

The phreatic groundwater level -a controlled *polder water level-* at a part of the *constant piezometric level* boundary is changed instantaneously. A sudden change in the level of the Haarlemmermeer polder is simulated: from -0.6 m *N.A.P.* (before 1854) to -5.5 m *N.A.P.* (after the reclamation in 1854). The changes in freshwater head are calculated with the adapted MOC for several values of specific storativity S_s at a observation well in the Haarlemmermeer polder area (x=7475 m and y=-105 m *N.A.P.*). As can be seen, fig. 9 shows that the new steady state situation is approached within half a year for all -realistic- specific storativity S_s. However, if the specific storativity S_s is much larger ($1 * 10^{-4} m^{-1}$) than the maximum observed value in the area of the Amsterdam Waterworks ($3 * 10^{-5} m^{-1}$), it takes about a one year before the new steady state situation is approached.

It is interesting to know that the duration of the flow time step is rather important, when the specific storativity S_s is not equal to zero. If the interest is focused on the development to a new steady state situation, the duration of the flow time step may not be too large, because the new steady state situation may be approached within some (tens of) days. In fig. 10, where the specific storativity $S_s = 3 * 10^{-5} m^{-1}$, the influence of the duration of the flow time step on the development of the freshwater head to a new steady state situation is shown. Simulations with large flow time steps generate inaccurate freshwater heads as a

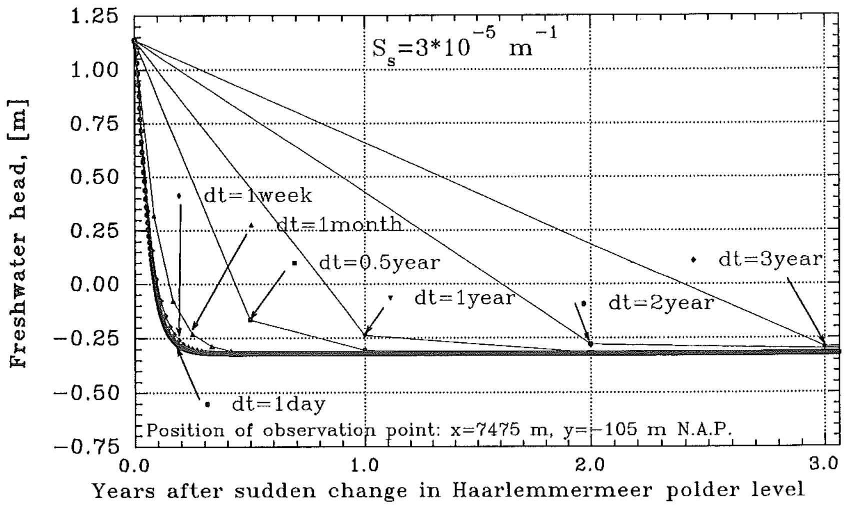

Figure 10: *The influence of the duration of the flow time step on the freshwater head as a function of about 3 years in a observation well with the position x=7475 m and y=-105 m N.A.P.. The specific storativity* $S_s = 3 * 10^{-5} m^{-1}$.

function of time. Fits with small flow time steps of days and weeks seems to approach the exact development of the freshwater head accurate enough. However, if only the first 8 days after the sudden reclamation of the Haarlemmermeer polder are considered (see fig. 11), fits of only hours seems to be accurate enough.

5.2 Influence of longitudinal dispersivity α_L

Fig. 12 gives the chloride distribution in the year 1988 for four different values of longitudinal dispersivity α_L: 0.02, 0.2, 2.0 and 20.0 m. The simulated and observed chloride distributions fit best, when a small longitudinal dispersivity α_L of 0.02 m is employed (Kooiman) [1]. The cases with small longitudinal dispersivities -α_L 0.02 m and 0.2 m- have a freshwater lens that correspond to observed values. The case with $\alpha_L = 2.0$ m shows a freshwater lens that is too small and the case with $\alpha_L = 20.0$ m does not simulate a freshwater lens any more: the geohydrologic system consists of only a large zone with brackish groundwater.

Fig. 13 shows the volume distributions in percentages of the geohydrologic system during the past 134 years for several longitudinal dispersivities. The groundwater volume is divided in a fresh ($Cl^- \leq 300$ mg/l), brackish and saline ($Cl^- \geq 10000$ mg/l) fraction. The geohydrologic system is bounded in the horizontal direction by x=0 m and x=12500 m, and in vertical direction by the levels y=-10 m and y=-170 m *N.A.P.* The figure shows that especially the case with $\alpha_L = 20.0$ m differs from the reference case which is the best fit. The fresh volume fraction shows that also the case with $\alpha_L = 2.0$ m deviates from the cases with relative small longitudinal dispersivities.

5.3 Influence of the loam layer

To assess the influence of the loam layer on the fresh, brackish and saline groundwater

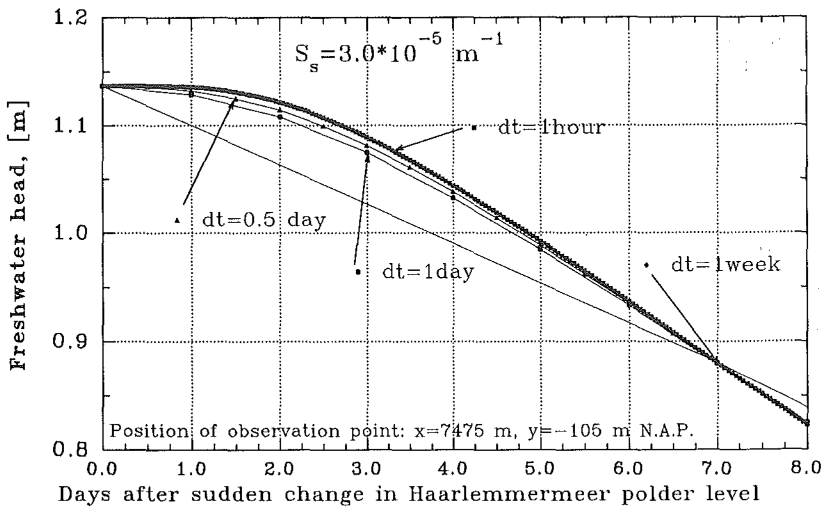

Figure 11: *The influence of the duration of the flow time step on the freshwater head as a function of the first 8 days in a observation well with the position x=7475 m and y=-105 m N.A.P.. The specific storativity* $S_s = 3 * 10^{-5} m^{-1}$.

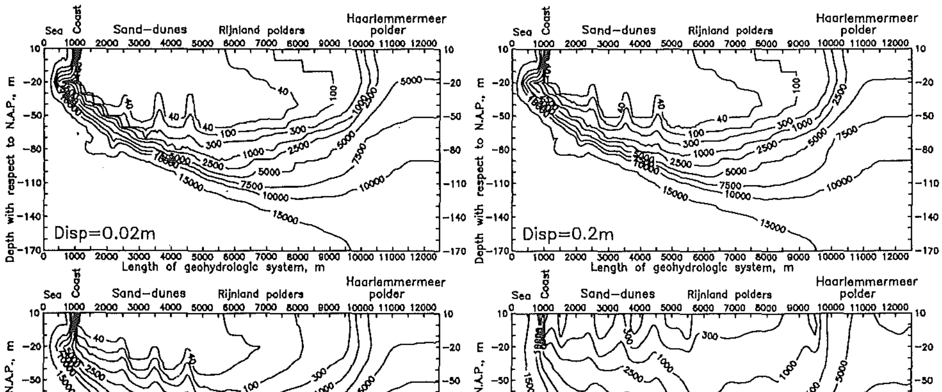

Figure 12: *The chloride distribution in the year 1988 for the reference case, calculated with four different longitudinal dispersivities* α_L*: 0.02 m; 0.2m; 2.0 m and 20.0 m.*

distribution the following cases, simulated for the next millennium, are compared: (a) the

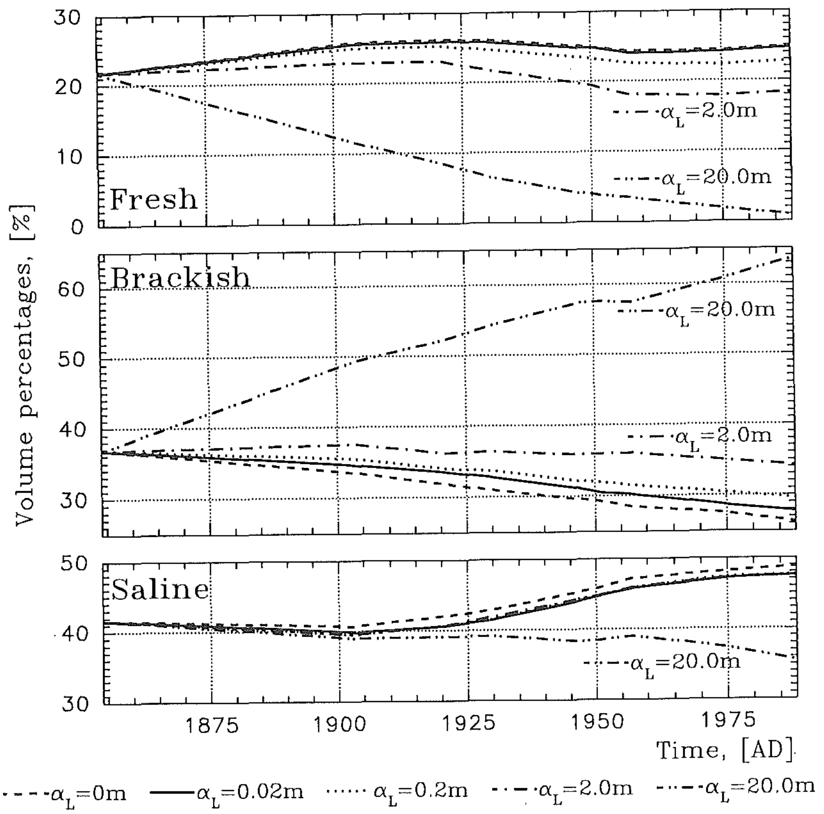

Figure 13: *The volume percentages of fresh, brackish and saline groundwater in the geohydrologic system during the past 134 years after the reclamation of the Haarlemmermeer for several longitudinal dispersivities.*

reference case; (b) a case with a sea level rise[6] of 60 cm per century[7]; (c) a case with no loam layer and no sea level rise; and (d) a case with no loam layer and a sea level rise of 60 cm per century. Since a long time period is simulated, many processes that affect the geohydrologic system should be taken into account, such as hydrological, morphological and man-induced processes. However, these processes are left out of consideration.

Fig. 14 gives the difference in chloride load[8] at a part of the Haarlemmermeer polder -between 9500 and 12500 m- for these four cases. Obviously, the presence of the loam layer keeps low the chloride load through the Holocene aquitard in the Haarlemmermeer polder.

[6]The present research is carried out in the framework of a project focused on impacts of changes in sea level.

[7]This global mean sea level rise is the best estimate under the Intergovernmental Panel on Climate Change (IPCC) Business-as-Usual emission scenario (1990). The sea level rise is imposed in the adapted MOC as 6-cm-elevations of freshwater head each decennium at the seaside boundary of the geohydrologic system.

[8]Chloride load through the Holocene aquitard is derived here from computed chloride concentrations and seepage rates.

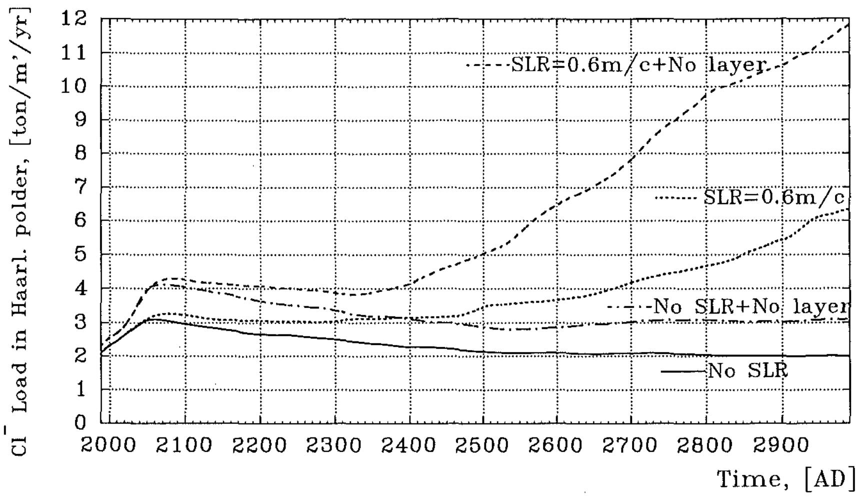

Figure 14: *The chloride load in ton per meter width per year at a part of the Haarlemmermeer polder. The influence of the loam layer between -70 and -90 m N.A.P. is visualized.*

In case of sea level rise, the influence of the loam layer is even greater.

Fig. 15 shows the volume distributions in percentages of the geohydrologic system for the next millennium for the four cases. The loam layer retards the salinisation process of the geohydrologic system. The geohydrologic system reacts faster on changes in boundary conditions when the loam layer is absent. For instance, in case of no loam layer, the freshwater lens -volume of fresh groundwater- would be enlarged more rapidly during the next three centuries than if the loam layer is present, nearly independent of sea level rise. However, after that period, -for the case with sea level rise- sea level has risen so much that the volume of fresh groundwater is decreasing quickly. And from the 25^{th} century on, the volume of fresh groundwater is smaller than of the case with sea level rise ánd loam layer.

CONCLUSIONS

The adapted two-dimensional vertical groundwater flow model MOC, which uses the 'particle tracking' method, can be used to simulate geohydrologic systems along the Dutch coast, where density flow and solute transport occur. Due to density flow, numerical instabilities -arising from the numerical approximation- may occur at places where the transition zone between fresh and saline groundwater is small and low vertical velocities occur. The numerical instability can be limited by adapting model parameters, such as to enlarge the number of particles per grid cells; to calculate the groundwater flow equation more often in relation to the time steps in which the particles are moved; or to calculate the groundwater flow equation more accurate. The influence of the number of particles per grid cell seems to be the greatest, followed by the size of the flow time step, and finally by the convergence criterion TOL. Furthermore, numerical instability can be counteracted if the initial density distribution is smoothened more, especially at transition zones. Enlarging the total number of grid cells does not necessarily decrease the numerical instability. For if grid cells are too small, a particle in an instability zone can easily enter a neighbouring grid cell which has a complete different density. Consequently, the numerical instability may be amplified.

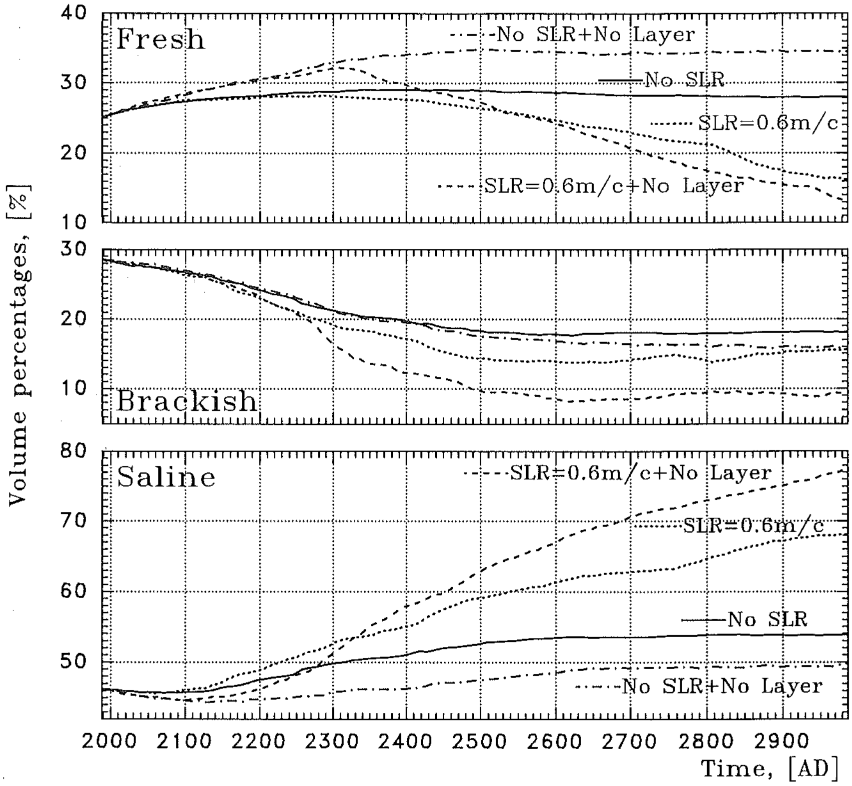

Figure 15: *The volume percentages of fresh, brackish and saline groundwater in the geohydrologic system during the next millennium for two cases with and two cases without a loam layer.*

Some geohydrological parameters have great influences on groundwater flow and solute transport in the chosen cross-section. If a time period in the order of only several weeks is considered in the simulation, the specific storativity S_s should not be taken equal to zero. Since the time lag, before a new steady state situation is approached, can not be neglected during that time period. However, normally much larger time periods are simulated with the adapted MOC, so in those cases the specific storativity S_s can be set to zero. In the chosen cross-section, dispersivities need to be relative small. The longitudinal dispersivity α_L must be set to a small value (in the order of centimeters or decimeters), so that calculated chloride distributions correspond to observations. Finally, the presence of the loam layer between -70 m and -90 m *N.A.P.* retards the salinisation process of this geohydrologic system significantly.

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