



Upward groundwater flow in boils as the dominant mechanism of salinization in deep polders, The Netherlands

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SUMMARY

As upward seepage of saline groundwater from the upper aquifer is leading to surface water salinization of deep polders in the Netherlands, we monitored the processes involved in the Noordplaspolder, a typical deep polder. Our results show three types of seepage: (1) diffuse seepage through the Holocene confining layer, (2) seepage through paleochannel belts in the Holocene layer, and (3) intense seepage via localized boils. They differ with regard to seepage flux, chloride concentration, and their location in the polder; thus, their contributions to surface water salinization also differ. Permeable, sandy paleochannel belts cut through the lower part of the Holocene layer, resulting in higher seepage fluxes than the diffuse seepage through the Holocene layer where there are no paleochannels. The average chloride concentration of paleochannel seepage is about 600 mg/l, which is sixfold higher than the average concentration of diffuse seepage. The highest seepage fluxes and chloride concentrations are found at boils, which are small vents in the Holocene layer through which groundwater preferentially discharges at high velocities. This results in upconing of deeper and more saline groundwater, which produces an average chloride concentration of 1100 mg/l. Despite the fact that seepage fluxes are difficult to measure, we were able to calculate that boils contribute more than 50% of the total chloride load entering the Noordplaspolder and they therefore form the dominant salinization pathway.

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1. Introduction

In many coastal areas, groundwater is saline because of sea water intrusion, marine transgressions and sea spray (Stuyfzand and Stuurman, 1994). The salinity of groundwater may pose problems for the sustainable exploitation of fresh groundwater for drinking water, irrigation or industrial purposes. Land subsidence, climate change and sea level rise accelerate salinization by enhancing the intrusion rate (Oude Essink, 2008), which means that low-lying deltaic areas, such as the deltas of the Mekong, Ganges, Mississippi, and Po rivers, are highly vulnerable (Barlow and Reichard, 2010; Bobba, 2002; Custodio, 2010; Giambastiani et al., 2007; Meisler et al., 1984; Ranjan et al., 2006). In the Netherlands, which has about 25% of its surface below sea level (Fig. 1), the upward seepage of saline and nutrient-rich groundwater into deep polders that have been reclaimed from lakes, leads to salinization and eutrophication of the regional surface water (Wesseling,

1980; Van Rees Vellinga et al., 1981; De Louw et al., 2000; Van Puijenbroek et al., 2004; Van den Eertwegh et al., 2006). This seepage makes the surface water unfit for irrigation and adversely affects aquatic ecosystems.

Since salinization by groundwater seepage has been a major water quality problem for a long time, there have been several studies on the water and chloride balances of deep polders in the coastal area of the Netherlands (ICW, 1976; Wit, 1974; Van Rees Vellinga et al., 1981; Pomper and Wesseling 1978; Griffioen et al., 2002). These earlier studies considered seepage through the confining top layer of clay and peat, with its low permeability, to be spatially uniform. However, this made it impossible to explain the high levels of salt load which were observed (e.g. ICW, 1976; Van Rees Vellinga et al., 1981). The earlier assumption of uniform upward seepage was obviously incorrect and indeed, many studies showed that groundwater seepage could occur through preferential pathways (e.g. Becker et al., 2004; Kalbus et al., 2009; Kishel and Gerla, 2002; LaSage et al., 2008). Preferential groundwater seepage may also have important implications for the chemical loading to surface waters (Keery et al., 2007; Tesoriero et al., 2009). The hydrochemical composition of preferential seepage can be used to distinguish this form of upward

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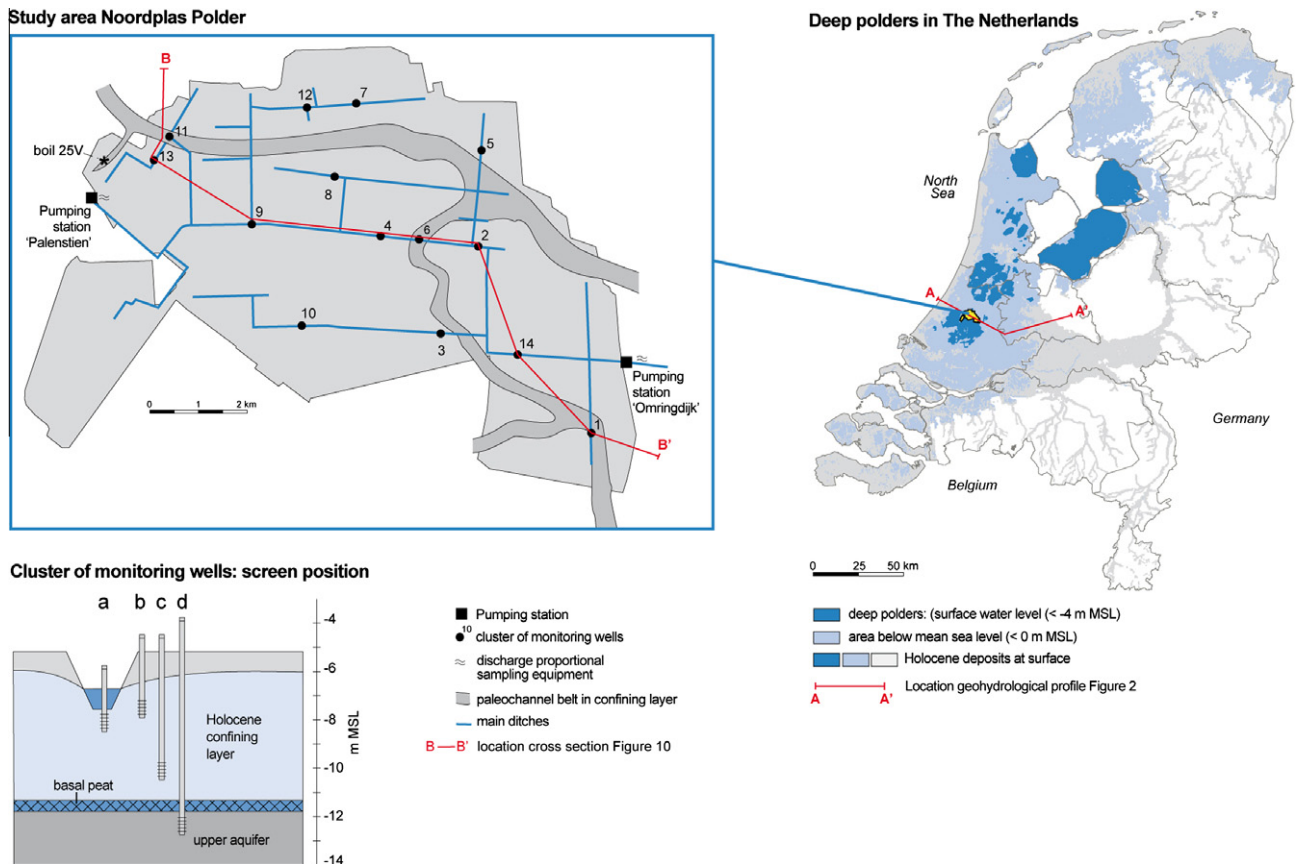


Fig. 1. Location of deep polders (reclaimed lake areas) in The Netherlands and of Noordplas Polder (study area), showing monitoring points, main ditches and the paleochannel belts in the Holocene confining layer.

groundwater flow from other types. Geological mapping of the lowlands of the Netherlands shows that permeable, sandy paleochannel belts partly cut through the confining top layer (Berendsen, 1998; Berendsen and Stouthamer, 2000; Weerts, 1996; Hijma et al., 2009) and thus act as preferential pathways for upward groundwater flow. An even more localized form of preferential groundwater flow, called boils, has been observed in deep polders (Vink, 1954; Rijnland, 1967, 1975, 1981; De Louw et al., 2004). Boils are vents that connect the underlying aquifer and the surface water or ground level through the confining top layer. They may develop when the pressure of upwelling water is greater than the pressure exerted by the weight of the overlying stratum. This produces heaving and cracking and the occurrence of pin boils, which may enlarge and become sand boils if the erosive forces increase (Holzer and Clark, 1993; Li et al., 1996; TACFD, 1999). Pin boils are small springs or upwellings of water, whereas sand boils are larger vents that discharge both water and sediments (Li et al., 1996). Permanent, large hydraulic gradients on the edges of deep polders in the Netherlands stimulate the development of such boils. Two types of boils have been described in the literature: *flood-induced boils* along artificial levees (Turnbull and Mansur, 1961; Kolb, 1976; Sellmeijer and Koenders, 1991; Holzer and Clark, 1993; Li et al., 1996; TACFD, 1999; Ozkan, 2003; Ojha et al., 2003), and *earthquake-induced liquefaction boils* (sand blows) (Seed and Idriss, 1967; Holzer and Clark, 1993; Obermeijer, 1995; Li et al., 1996, 2004). However, because boils are a fairly common feature of polder areas, and perhaps in other areas with relatively large, vertical pressure gradients, they have been given little attention in the deep polders. The contribution of boils and seepage through paleochannel belts to the salinization of deep polders is therefore unknown.

Here we present evidence for different types of upward groundwater seepage in a deep polder and, from independent physical and chemical observations, we deduce how much boils contribute to chloride loads in the surface water of the polder.

2. Material and methods

2.1. Study area

The study area is a typical deep polder in the west of the Netherlands called the Noordplas Polder (Fig. 1). It covers 37 km², of which 86% is used for agriculture (arable farming 62%; pasture 24%) and 14% is urban area. This former lake was reclaimed in different stages between 1750 and 1850 AD (Schultz, 1992). Its average soil surface level is –5 m below mean sea level (MSL) which is 2.5 m lower than the surrounding peat lands (Fig. 2). The polder is drained by an artificial network of ditches and underground drains. The drained water is pumped out of the polder into a higher-lying canal at two pumping stations and then conveyed to the sea. During dry periods, fresh water from the regional surface water network is admitted to the polder via inlets (inlet water) to maintain the proper level of surface water and to decrease the chloride concentrations in the surface water by flushing (Van de Ven, 2003). During these dry periods, both the inflow of water via inlets and the outflow via pumps occur simultaneously. The surface water levels in the polder ditches are kept at a nearly constant level of about –6.2 m MSL.

The geohydrological situation is characterized by two (semi-) confined aquifers of Pleistocene sands, as shown in Fig. 2. The upper aquifer is covered by a 6–9 m thick, confining layer of

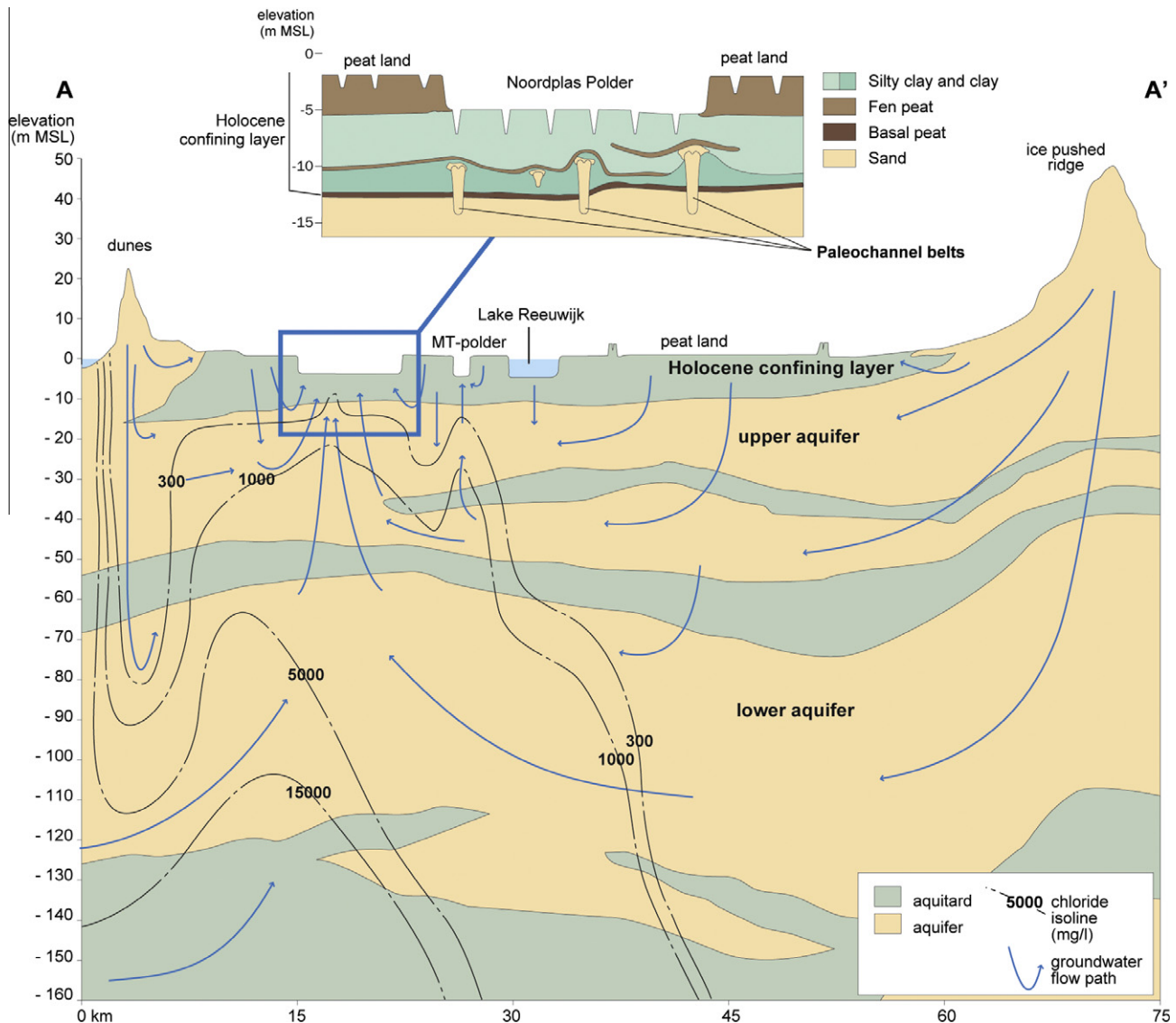


Fig. 2. Regional geohydrological profile across the study area, perpendicular to the North Sea. The position of the aquifers and aquitards is based on descriptions from boreholes at a density of $0.25/\text{km}^2$ (REGIS II, 2005). The chloride distribution is derived from the Dutch national database (www.dinoloket.nl). Position of profile A–A' is indicated in Fig. 1.

Holocene peat, loam and clay of low permeability (Mulder et al., 2003). The hydraulic conductivity of these sediments varies between 10^{-4} m/d and 10^{-1} m/d (Weerts, 1996). The lower part of this Holocene layer is the less permeable part which consists of a 10–50 cm thick compacted, mineralized peat (basal peat). At various locations, the compacted peat and overlying loam and clay deposits have been eroded by streams and tidal channels during the Holocene (Van der Valk, 1996; Hijma et al., 2009). We mapped the locations of these sandy paleochannel belts in the Noordplaspolder (Fig. 1) by using the lithological borehole descriptions of the Holocene confining layer derived from the Geological Survey of the Netherlands' database (www.dinoloket.nl). We found the channels are about 200–300 m wide, at a depth of 3–5 m below the surface. They are in direct contact with the upper aquifer and act as preferential pathways in the Holocene confining layer because of their higher permeability (Fig. 2). Weerts (1996) found hydraulic conductivities in the range of 1.0–10.0 m/d for these sandy paleochannel deposits.

The groundwater in the Pleistocene aquifers is brackish to saline and shows an increasing trend of chloride concentrations with

depth (Fig. 2) (Van Rees Vellinga et al., 1981; Oude Essink, 1996). The origin of salt in this groundwater is mainly linked to the Holocene transgressions (Volker, 1961; Pomper and Wesseling, 1978; Stuyfzand, 1993; Post, 2004; Stuyfzand and Stuurman, 2008). Hydraulic heads in the upper aquifer exceed surface water levels in the polder, resulting in upward seepage of brackish-to-saline groundwater from the upper aquifer into the overlying Holocene layer (Van Rees Vellinga et al., 1981; De Louw et al., 2000). This seepage water finally discharges into the surface water system.

2.2. Monitoring program

The groundwater and surface water monitoring program in the Noordplaspolder aimed to identify and quantify water fluxes and chloride loads into the surface water. To determine the daily polder water discharge, the pumping time was recorded automatically at the two pumping stations (Fig. 1). Multiplying the total daily pumping time by the pumping rate gave the daily volume of polder water discharge. The pumping water was sampled automatically with a frequency proportional to the discharge, to determine the

chloride loads that leave the polder. The samples were mixed automatically into one bulk sample and collected every 3–7 days; they were analyzed in the laboratory for their chloride content. Multiplying the chloride concentration by the volume of polder water discharged gave the total chloride load being pumped out of the polder. We sampled the inlet water at 5 major inlets and analyzed it for chloride on a monthly basis. Precipitation was measured at two locations in the polder with a tipping bucket rain gauge. The chloride concentration of precipitation was measured once. Daily sums of the potential reference crop evapotranspiration (Makkink, 1957) were obtained from two nearby meteorological stations (Valkenburg and Schiphol) of the Royal Netherlands Meteorological Institute (KNMI).

We installed 14 clusters of groundwater monitoring wells to quantify the water levels and to analyze water composition. The numbered monitoring clusters are shown in Fig. 1. We installed 3 clusters (1, 6 and 11) in areas with paleochannel belts in the Holocene confining layer and 11 clusters outside these areas. Each cluster contained 4 monitoring wells with screens at different depths to measure the change of heads and groundwater composition with depth (Fig. 1). The 0.5-m-long screens of the monitoring wells are located just below the ditch bottom (a), in the Holocene confining layer at 2–3 m (b), and at 4–5 m depth (c), and in the upper aquifer at 6–8 m depth (d). Heads were measured every two weeks and 30 monitoring wells were automatically logged every hour. The heads were corrected for density differences by conversion into fresh water heads (Post et al., 2007). At every monitoring cluster, we installed a gauge in the ditch for surface water level observations. We sampled and analyzed the surface water and groundwater from screens a, c and d of each cluster twice (April 1999 and September 1999). Before taking groundwater samples, we extracted water from the monitoring wells until pH and EC became stable and the extracted volume was at least three times the volume of the monitoring wells. In the field, we measured pH, temperature (T), electrical conductivity (EC), dissolved oxygen (DO) and bicarbonate (HCO_3). Samples for cation analysis were filtered over 0.45 μm and preserved with 1% concentrated suprapure HNO_3 . In the laboratory, we measured major ions (Na, K, Ca, Mg, Cl, and SO_4), nutrients (NH_4 , NO_3 , NO_2 , N-kjel, ortho-P, tot-P) and iron (Fe). We used Stuyfzand's method (1989) to classify the groundwater into chemical types.

We mapped the chloride concentration of the surface water in the main ditches of the polder to determine the spatial variation of saline seepage. To obtain large contrasts in the surface water salinity, the survey was carried out in a dry period (August 1999) when dilution of the seepage water by rainfall was minimal. We took samples from the surface water at 250 m intervals and analyzed them for chloride in the laboratory.

A first attempt to map the boils and to determine their chloride concentration in the Noordplaspolder was done by the Rijnland Water Board in 1967 (Rijnland, 1967). In our field work, we updated their map on the basis of the boils' visible properties and took samples of the boil water for chloride analysis. Boils are evident as small vents in ditches (Fig. 3a), where water flows out or where "boiling" of quicksand occurs, as small sand volcanoes (Fig. 3b and d), or as collapsed ditch banks (Fig. 3e). We observed methane gas escaping through boils (Fig. 3c). Boils may be supported by several interconnected conduits in the Holocene confining layer originating from one source vent in the upper aquifer (Fig. 3a, g, and h). For 15 of the mapped boils, we measured their upward seepage flux and estimated the upward flow velocity. Upward seepage fluxes were determined by collecting the outflowing boil water in a measuring cup during a certain time span. Where the outflow occurred underwater, instead of on land or in ditch banks, we isolated the outflowing boil water from ditch water by using a tube and then measured its flux. Upward flow velocities

were estimated by dividing the measured seepage flux by the cross-sectional area of the outflow vent.

We studied one boil in greater detail (boil 25 V in Fig. 1). We installed 6 monitoring wells at 0, 5, 10, 25, 30, 75 m from this boil, with their screens located in the upper aquifer (7 m depth). Water samples were taken from these wells, from surface water and from the boil water, and analyzed for chloride content. We measured the soil temperature in and around the boil to trace lateral variations in upward groundwater flow and to evaluate its flow pattern (Becker et al., 2004; Taniguchi et al., 2003; Bense and Kooi, 2004; Van Wirdum, 1991; Cartwright, 1974). The soil temperature was measured with a temperature probe in and around the boil, at up to a depth of 4 m and at intervals of 0.1 m.

3. Results

3.1. Heads

In Fig. 4 we show the observed heads for groundwater monitoring wells in cluster 3. Hydraulic heads in the upper aquifer (screen d) exceed both the surface water levels as well as the heads in the Holocene confining layer (screens a–c), indicating a permanent upward groundwater flow from the upper aquifer. Averaged over all the monitoring clusters, the head difference between the upper aquifer and Holocene layer was as much as 1.0 m. These head differences, and consequently seepage fluxes, do not fluctuate much throughout the year. The maximum yearly amplitude of the head differences was less than 20 cm. These observations for cluster 3 are in full agreement with those for all 13 monitoring clusters.

3.2. Polder water discharge and chloride concentration

The daily discharge and chloride concentration of the water pumped out of the polder is shown in Fig. 5. The total polder water discharge was equal to 33.8 $\text{M m}^3/\text{y}$, whereas the measured precipitation surplus is only 19.5 $\text{M m}^3/\text{y}$. The large difference between the polder water discharge and the precipitation surplus indicates that other sources must contribute significantly to the water input, i.e. upward groundwater seepage from the upper aquifer and the admission of inlet water. The total measured chloride load pumped out of the polder equals 11.6 M kg/y . Periods of higher discharge are typically characterized by low chloride concentrations because of rainwater dilution. These low concentrations (<350 mg/l) occur in winter and during wet periods in summer. During dry periods, the relative contribution of saline seepage increases and chloride concentrations reached up to 720 mg/l. The yearly average chloride concentration of the polder water discharge is about 400 mg/l.

The surface water chloride concentration in August 1999 varied spatially between 50 mg/l and 2200 mg/l (Fig. 6). We found elevated chloride concentrations (>600 mg/l) near paleochannel belts and at boils.

3.3. Groundwater composition

In Table 1 we show the average concentrations of the main constituents of the surface water and groundwater at different depths, for the 3 monitoring clusters in areas with paleochannel belts in the confining layer, and the 11 clusters in areas without paleochannel belts. The differences in concentrations between the analyzed groundwater in April 1999 and September 1999 are small (<5%). However, the chloride concentration of groundwater measured in the upper aquifer directly below the Holocene confining layer (screen d) is spatially quite variable, ranging from 33 mg/l to 675 mg/l (Fig. 6). The average chloride concentration in the upper aquifer for areas with paleochannel belts was

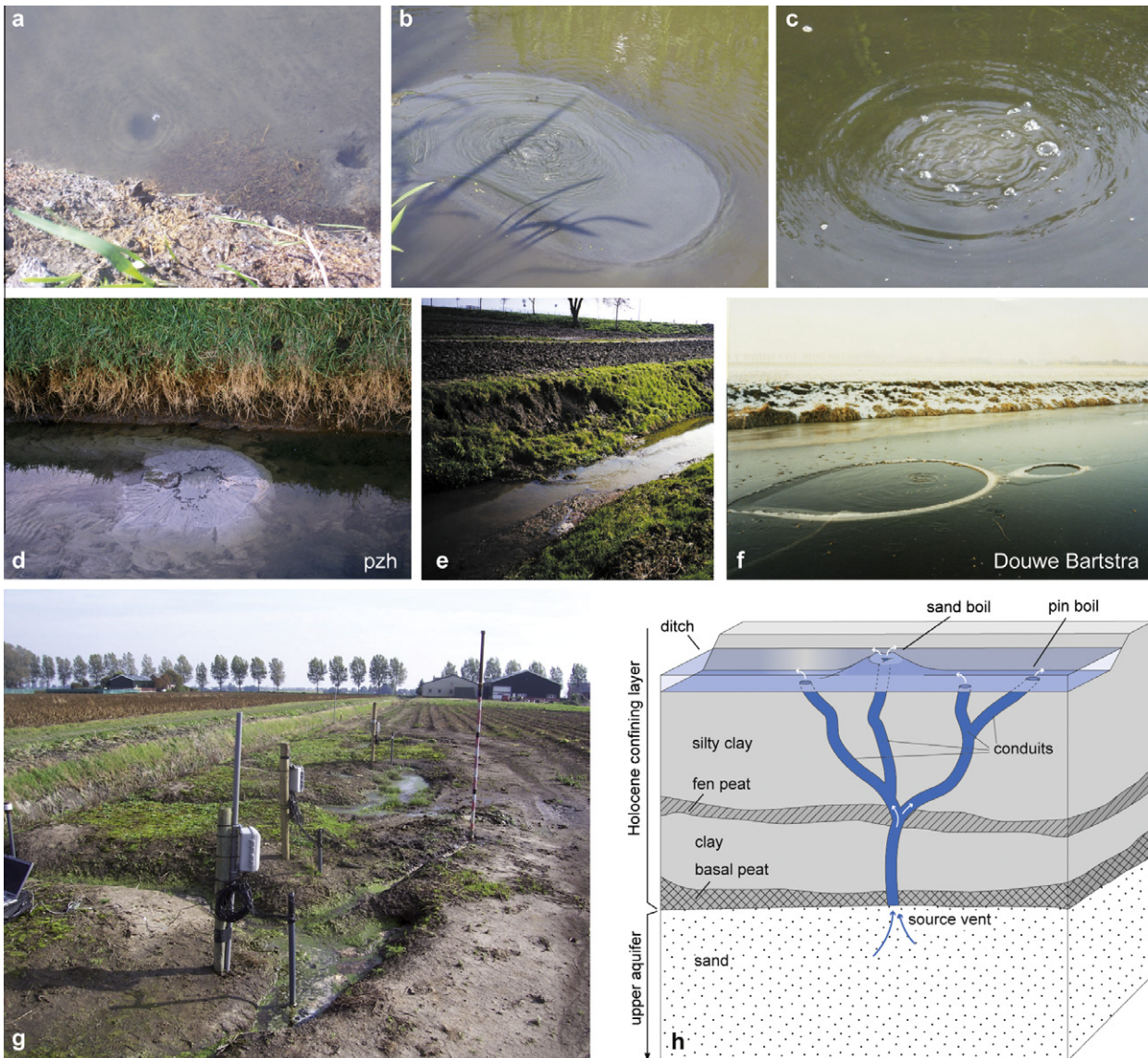


Fig. 3. Boils in deep polders: (a) pin boil, (b) sand boil, (c) a boil emitting methane, (d) sand volcano, (e) collapsed ditch bank, (f) hole in ice caused by warmer water welling up from boil, (g) sand boils on land, (h) schematic diagram of boils with several conduits.

592 mg/l. In the 11 sites without paleochannel belts, the average chloride concentration was much lower at 117 mg/l (Table 1). Another remarkable difference are the high SO_4^{2-} concentrations in the Holocene layer (screens a and c) for the areas without paleochannel belts, whereas concentrations of SO_4^{2-} are consistently and significantly smaller in areas with paleochannel belts (Table 1).

3.4. Boils

At 54 locations in the Noordplaspolder we observed intense upward groundwater flow in boils (Fig. 6); 17 of these were mapped in 1967 (Rijnland, 1967). We found that 85% of the boils occur in ditches or ditch banks, and only 15% occur on land. As expected, most visible boils occur at the edges of the polder, in ditches and on top of paleochannel belts (Fig. 6). At the edges of the polder, the heads in the upper aquifer are largest, increasing the probability of boil development. Underneath ditches and at locations with paleochannel belts, the weight of the Holocene layer is markedly reduced, which also significantly increases the risk of boil develop-

ment. For 49 of the observed boils it was possible to determine their chloride concentration, from which the average chloride concentration of the boils was found to be as much as 1100 mg/l, with a median of 1300 mg/l. The larger boils tended to have even higher concentrations, with a recorded maximum of 2850 mg/l (Table 2). Fluxes measured at 15 individual boils ranged from 0.5 m³/d to 100 m³/d and estimated flow velocities attained as much as 800 m/d. It must be noted that it was not possible to map all the boils, because they are not visible in the main water courses, which are too wide, too deep and too turbid due to eutrophication. Therefore, it was also not possible to upscale the measured boil fluxes to a total boil flux for the whole polder.

Fig. 7 shows the results of detailed soil temperature–depth measurements in and around boil 25 V at the end of the summer period (1999). The soil temperature in the boil was relative constant with depth (about 11 °C), whereas outside the boil the soil temperature was much higher and decreased from about 17 °C at surface level to about 13 °C at 3 m depth. Detailed measurements around the same boil showed that the chloride concentration of boil water was nearly twice as high as that of groundwater in

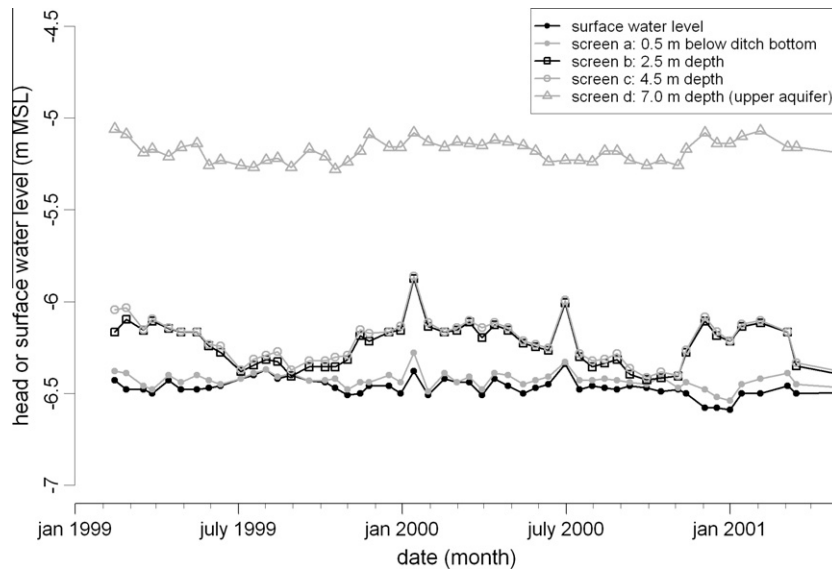


Fig. 4. Measured surface water level and hydraulic heads in Holocene confining layer (screens a–c) and upper aquifer (screen d) for monitoring wells in cluster 3.

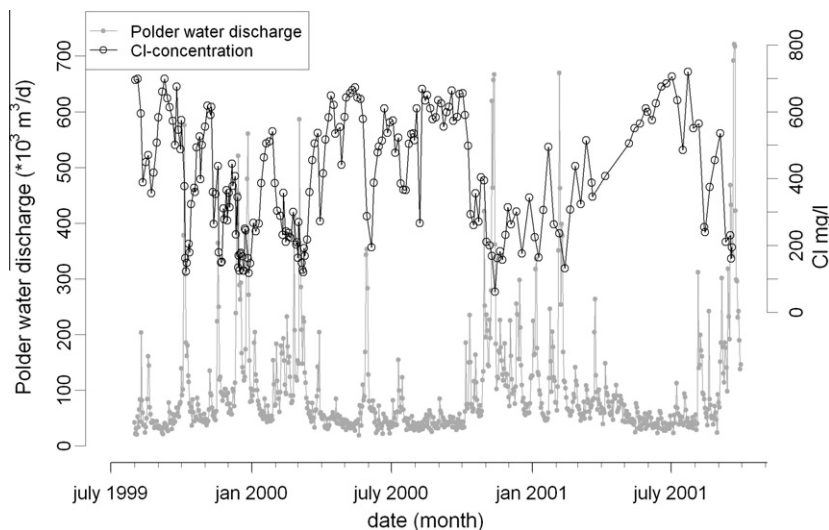


Fig. 5. Daily discharge volume and chloride concentration of water pumped out of the polder over period July 1999–September 2001.

the upper aquifer at 7 m depth (Fig. 8). The chloride concentration of the surface water downstream of the boil was about twice as high as upstream of the boil.

4. Discussion

4.1. Types of seepage

The observed head losses in the upward direction were persistent in space and time (Fig. 4) and show that there is permanent upward groundwater seepage from the upper aquifer into the overlying Holocene confining layer for the entire area that was monitored. The large spatial variation of chloride concentration of the surface water (Fig. 6) implies that upward saline seepage is not uniform in the polder. The elevated chloride concentrations (>600 mg/l) of the surface water at boils and near areas with paleochannel belts in the Holocene layer indicate that upward saline seepage occurs preferentially at these locations. These paleochannel belts may well form a conduit for upward groundwater flow

because they consist of permeable sands and cut through the lower part of the Holocene layer, which is less permeable. The locations of these paleochannel belts could be mapped relatively accurately using borehole descriptions of the Holocene layer (about 10 borehole records per km², www.dinoloket.nl).

Based on these findings, we have been able to distinguish three types of upward groundwater seepage: (1) diffuse seepage, (2) paleochannel seepage, and (3) intense seepage via localized boils (Fig. 9). Diffuse seepage is seen over the majority of the polder (31 km²), with groundwater flowing upward through the Holocene sediments (peat, loam, clay) of low permeability. Paleochannel seepage is associated with the much more permeable, sandy, paleochannel belt deposits in the Holocene layer and is seen only in the area with these belts (6 km²). Upward groundwater flow fluxes through these permeable, sandy paleochannel belts are higher than the diffuse seepage fluxes because of the hydraulic conductivity differences. Boils are small, localized vents in the Holocene layer through which groundwater is discharged at high velocities.

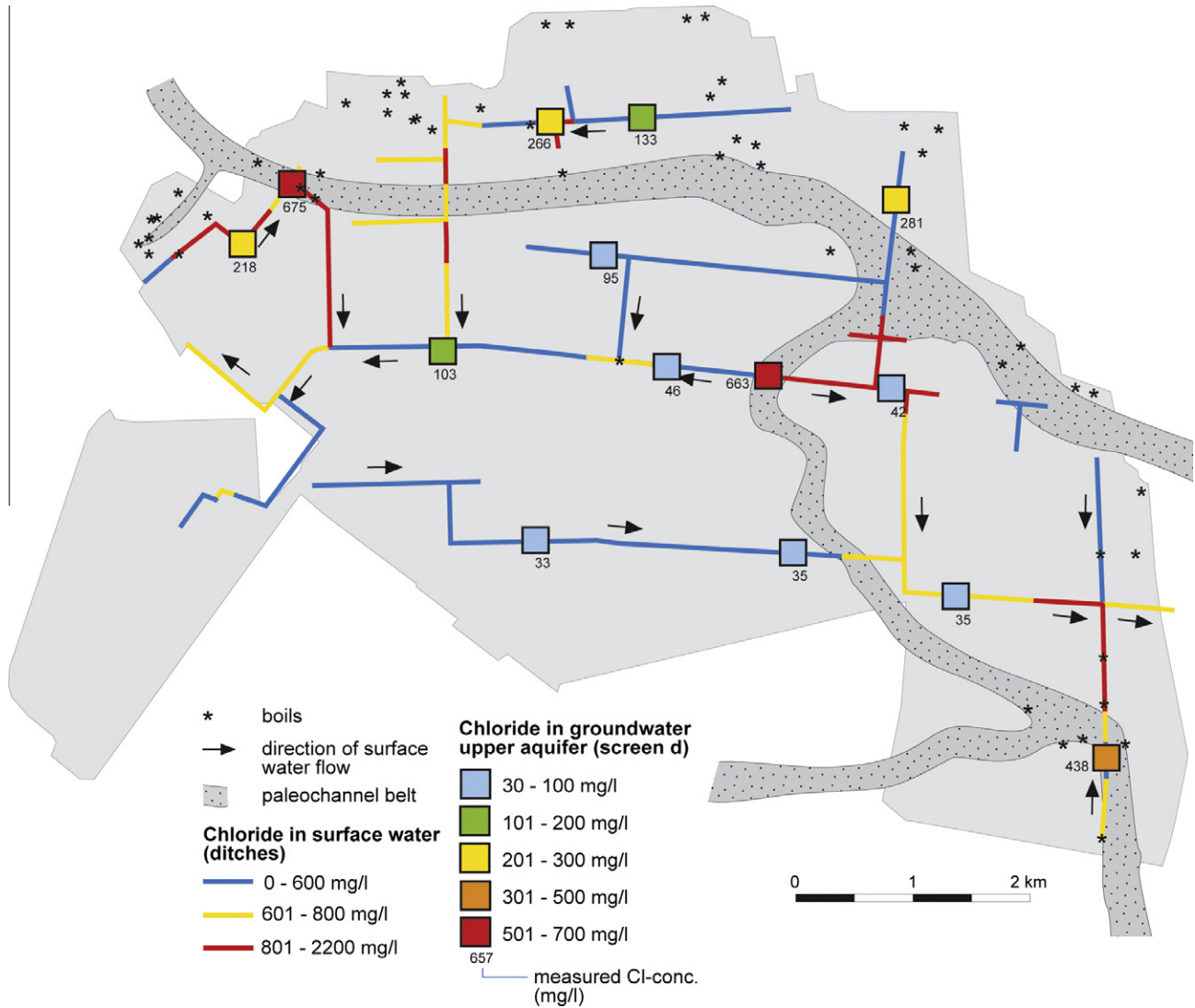


Fig. 6. Chloride monitoring results (mg/l) for the Noordplas Polder: surface water in August 1999 (sampling interval 250 m) and groundwater at the top of the upper aquifer, screen d (average of April and September 1999). The mapped boils and location of the paleochannel belts in the Holocene confining layer are indicated.

Table 1
Average concentrations of the main chemical constituents (in mg/l), electrical conductivity (EC, mS/cm) and pH (-) of surface water and groundwater at different depths (screens a, c, and d) for areas with and without paleochannel belts in the Holocene confining layer (HCl).

No paleochannel belts in HCl (N = 11, Freq = 2)	Cl	NH ₄	NO ₃	HCO ₃	SO ₄	Ortho-P	Total-P	K	Mg	Na	Ca	Fe	EC	pH
Surface water	267	2.8	1.8	528	312	0.1	0.4	15	35	146	251	4.0	n.d.	n.d.
Below ditch bottom (a)	186	17.6	1.0	719	419	1.0	2.0	24	54	139	287	4.2	2.0	6.58
Confining layer, 5 m (c)	117	35.8	0.2	1046	250	4.3	5.2	27	82	132	202	2.1	2.0	6.57
Upper aquifer (d)	117	37.3	0.2	1024	12	1.3	2.9	25	67	119	114	13.8	1.7	6.46
Paleochannel belts in HCl (N = 3, Freq = 2)														
Surface water	608	5.6	1.0	580	270	<0.05	0.2	15	47	283	283	5.1	n.d.	n.d.
Below ditch bottom (a)	1245	18.7	<0.05	1050	8	<0.05	0.4	19	77	718	236	12.8	4.7	6.49
Confining layer, 5 m (c)	376	28.5	<0.05	976	17	1.9	2.6	21	69	247	185	18.8	2.4	6.43
Upper aquifer (d)	592	27.5	<0.05	1127	9	<0.05	1.1	23	73	369	194	26.8	3.1	6.45

N = number of locations, Freq = number of measurements per location, n.d. = not determined.

4.2. Seepage fluxes

Besides the observations of intense outflow of water and the high flow velocities recorded at the boils, there is another quite different indicator that strongly supports the preferentiality of flow and transport in boils. Temperature profiles around boil 25 V clearly showed a flow pattern with very localized, rapid, vertical

upward flow (Fig. 7). In the Netherlands, seasonal variations in atmospheric temperature normally penetrate into the soil to a depth of 15 m at most (Bense and Kooi, 2004). The year-round soil temperature at this depth is 10.5 °C to 11 °C. At boils, the penetration of atmospheric temperatures into the soil is counteracted by the rapid upward flow of groundwater with a constant temperature of about 11 °C (since it comes from deeper strata). This results

Table 2

Chloride concentrations (mg/l) of diffuse seepage, paleochannel seepage, boils, polder water discharge, inlet water and precipitation in the Noordplas Polder for monitoring period 01-08-1999 up to 01-08-2001.

	Mean	Median	Sdev	Min	Max	N	Freq
Diffuse seepage	117	95	96	33	281	11	2
Paleochannel seepage	592	663	134	438	675	3	2
Boils	1100	1300	638	337	2850	49	1
Polder water discharge	404	392	177	61	720	2	172
Inlet water	223	135	147	115	455	5	12
Precipitation	12	12	-	12	12	1	1

Sdev = standard deviation, N = number of locations, Freq = number of measurements per location.

in shallow, constant temperature profiles, in contrast with the rest of the area (Fig. 7).

Although fluxes at boils can be measured, there is a general consensus that it is extremely difficult to quantify the fluxes of paleochannel and diffuse seepage (e.g. Keery et al., 2007; Surridge et al., 2005; Murdoch and Kelly, 2003; Kishel and Gerla, 2002). Some of the underlying reasons are that they can be estimated using Darcy's equation but the vertical hydraulic conductivity of the

Holocene layer is very uncertain (Weerts, 1996), as well as being highly variable in space. However, our chemical groundwater data demonstrate that seepage fluxes must be much higher in the areas with paleochannel belts cutting into the Holocene layer. To illustrate this, the classified chemical groundwater types (screens a, c and d) were put on a cross-section together with geological data, hydraulic heads and deduced flow paths (Fig. 10). The dominant anion in the Holocene layer for the areas without paleochannel belts is SO_4^{2-} , whereas the dominant anion is Cl^- and concentrations of SO_4^{2-} are low in the areas with paleochannel belts and in the upper aquifer (see also Table 1). The elevated concentrations in the Holocene layer without paleochannel belts, as compared to the low concentrations in the upper aquifer, indicate that SO_4^{2-} must be enriched in the Holocene layer. The sources of the sulphates are thought to be pyrite (FeS_2), which is abundant in the Holocene organic clays, and peat layers, which release SO_4^{2-} upon oxidation (Ritsema and Groenenberg, 1993; Dellwig et al., 2001; Table 3). For this to occur, oxic- or nitrate-containing groundwater derived from the surface has to locally penetrate the confining layers down to 5 m depth before flowing upwards again to the surface (Fig. 10). Along these flow paths, acid produced during pyrite oxidation will dissolve calcium carbonates (Table 3) that are also

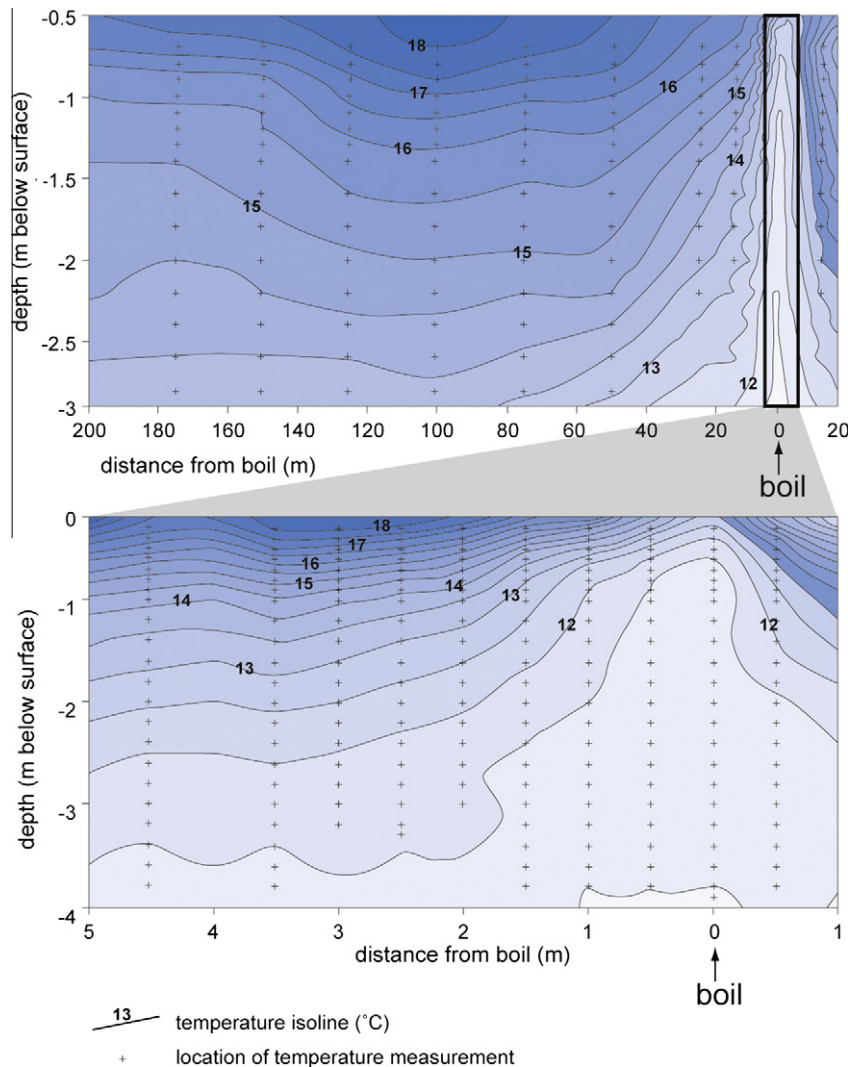


Fig. 7. Temperature–depth profiles in and around sand boil 25 V in the Noordplas Polder (see Fig. 1 for location). Crosses indicate the depth of temperature measurements (on 15 September 2006).

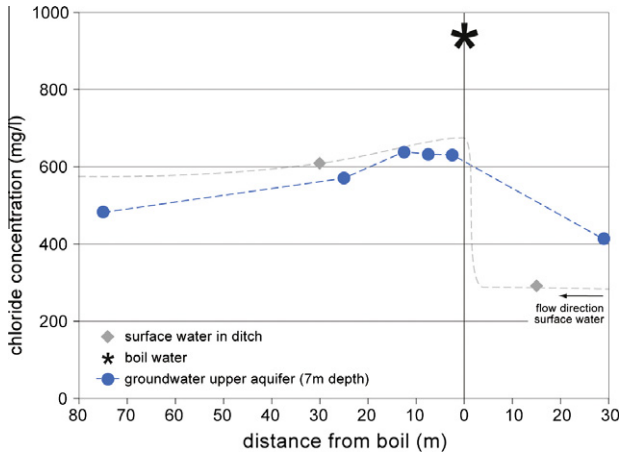


Fig. 8. Chloride concentrations of boil water, of surface water in ditch upstream and downstream from boil 25 V, and of groundwater at the top of the upper aquifer at various distances from the boil.

the permanent upward flow from the upper aquifer into the Holocene layer, these lenses of locally infiltrated groundwater can only reach depths of 4–5 m depth when upward groundwater fluxes from the upper aquifer are small. Our head measurements confirmed this. The largest head losses (on average 1.0 m) were found between the upper aquifer at 6–7 m depth and the Holocene layer at 4–5 m depth (Fig. 4). This suggests that the lowest permeability is found at the base of the Holocene layer, which is probably caused by the presence of basal peat. We would argue that the lower part of the Holocene layer is locally quasi-impermeable, so that no groundwater from the upper aquifer can flow up into it in some areas. The paleochannel belts with their high hydraulic conductivities cut through this lower, less permeable part of the Holocene layer, and allow for greater upward groundwater fluxes (Fig. 10). This relatively strong upward flow prevents shallow groundwater containing oxidants from reaching depths of 4–5 m. Groundwater of the same type (b-NaCl), originating from the upper aquifer, is therefore found in the entire Holocene layer above the paleochannel belts (Fig. 10).

abundant in the Holocene layer which contains sediments of marine origin (Ritsema and Groeneweg, 1993). Therefore, f-CaSO₄ groundwater types are found throughout the Holocene confining layer except in areas with paleochannel belts (Fig. 10). In view of

4.3. Salt water upconing

Diffuse-, paleochannel-, and boil seepage all originate from the upper aquifer (Fig. 9). The composition of the groundwater in the upper aquifer directly under the Holocene confining layer

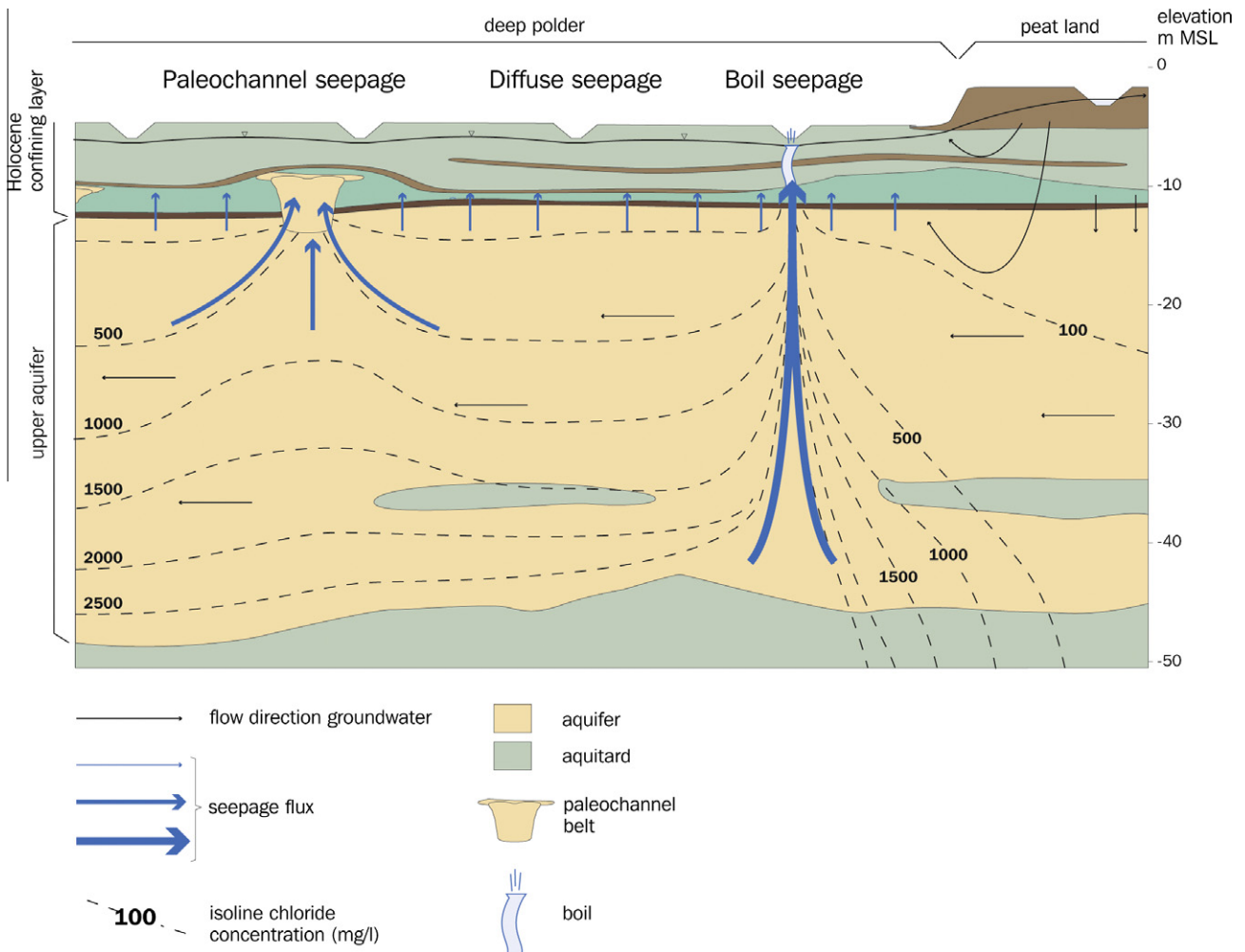


Fig. 9. Diagram showing upconing mechanism for the three seepage types with different fluxes and chloride concentrations: diffuse seepage, paleochannel seepage and boil seepage.

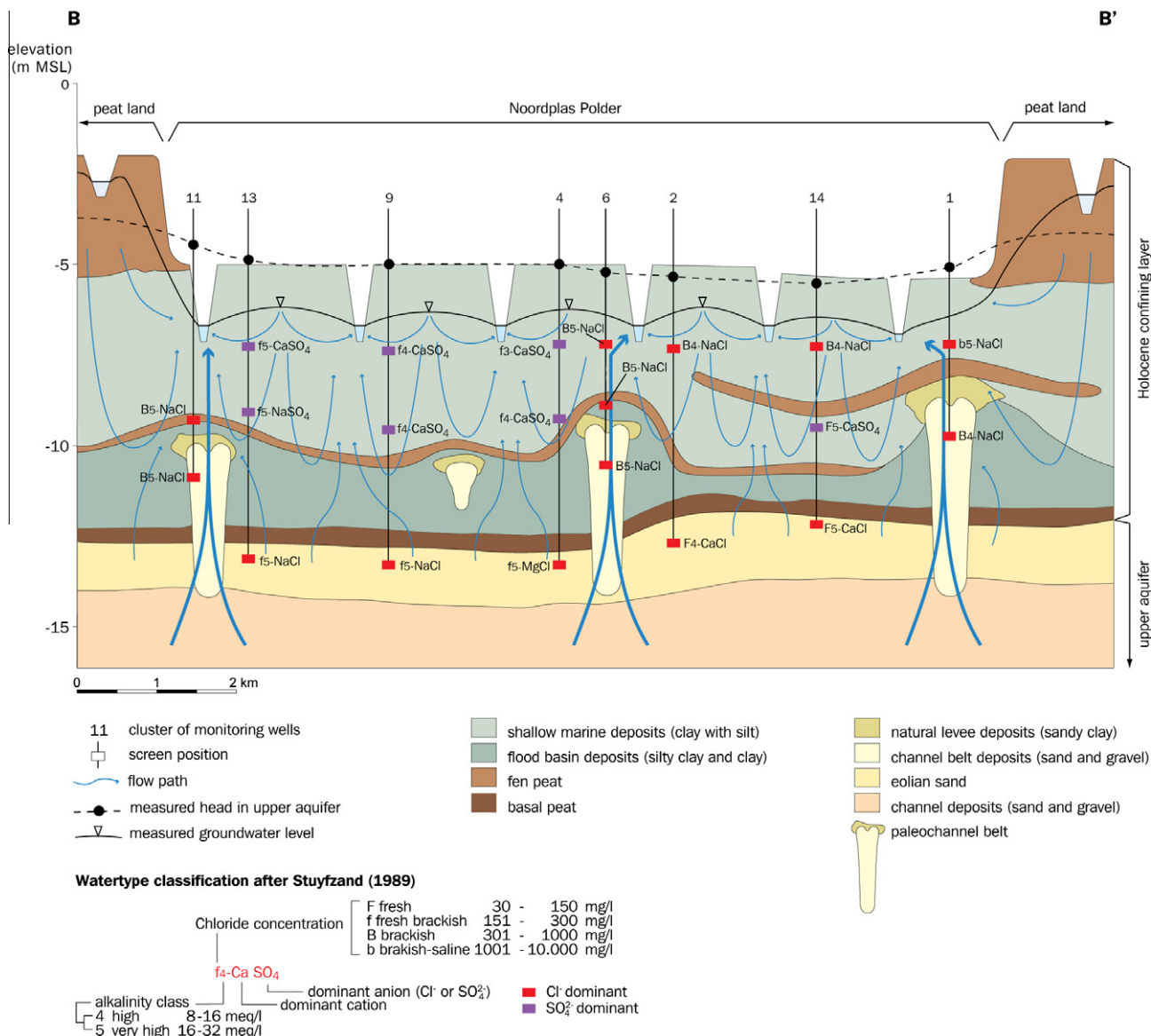


Fig. 10. Cross-section across Noordplas Polder with groundwater flow patterns in the Holocene confining layer based on different water types, measured hydraulic heads, and the sedimentary structure of the Holocene confining layer. Position of cross section B–B' is indicated in Fig. 1.

Table 3
Possible chemical reactions along flow path of local groundwater in the Holocene confining layer.

Description	Chemical reaction equation
Oxidation of pyrite by oxygen	$FeS_2 + 7/2 O_2 + H_2O \rightarrow Fe^{2+} + 2SO_4^{2-} + 2H^+$
Further oxidation of Fe ²⁺ by oxygen	$Fe^{2+} + 1/4 O_2 + 5/2 H_2O \rightarrow Fe(OH)_3 + 2H^+$
Oxidation of pyrite by nitrate	$FeS_2 + 14/5NO_3^- + 4/5H^+ \rightarrow Fe^{2+} + 2SO_4^{2-} + 7/5N_2 + 2/5H_2O$
Further oxidation of Fe ²⁺ by nitrate	$Fe^{2+} + 1/5NO_3^- + 12/5H_2O \rightarrow Fe(OH)_3 + 1/10N_2 + 9/5H^+$
Dissolution of calcium carbonates	$CaCO_3 + H^+ \rightarrow Ca^{2+} + HCO_3^-$

therefore determines that of the seepage water. There is a remarkable and consistent difference between the chloride concentrations of the three seepage types. For diffuse seepage, we found low concentrations of about 100 mg/l and for paleochannel seepage, we found higher chloride concentrations of about 600 mg/l (Table 2). Boils show both significantly and consistently higher chloride concentrations (average about 1100 mg/l) than the diffuse and paleochannel seepage waters. Hence, relatively high chloride concentrations in the upper aquifer occur consistently at those places

where preferential flow occurs. This leads us to believe that the higher fluxes at these preferential conduits are the reason for higher chloride concentrations, as higher fluxes lead to upconing of deeper and more saline water. It has been well established that chloride concentrations of groundwater in the aquifers below Dutch polders increase with depth, as shown in Fig. 2 (ICW, 1976; Pomper and Wesseling, 1978; Van Rees Vellinga et al., 1981; Griffioen et al., 2002; Post, 2004). The upconing mechanism of the three seepage types with their different upward

groundwater fluxes leading to different chloride concentrations is shown in Fig. 9. The mechanism is analogous to salt water upconing in response to pumping of groundwater, which has been described by e.g. Reilly and Goodman (1987), Ma et al. (1997) and Bower et al. (1999). They show that the salinity and depth of the pumped groundwater increase with pumping rate. Tellam et al. (1986) showed regional upconing of saline groundwater in response to low heads at springs in the Mersey Valley, UK, similar to the upconing mechanism in the Noordplaspolder.

Direct evidence of such salt water upconing by preferential seepage is provided by our chloride measurements around boil 25 V (Fig. 8). As all seepage originates from the upper aquifer, the large differences in concentration over such a short distance can only be explained by local upconing of deeper and therefore more saline groundwater as a result of high upward flow velocities.

4.4. Chloride loads from boils

Groundwater flowing up from the upper aquifer, either via the Holocene confining layer, through paleochannel belts or through boils, discharges into the ditches. After mixing with both inlet water and rainwater, it is pumped out of the polder. The daily discharge and chloride concentration of the water pumped out, as shown in Fig. 5, is therefore a mix of all these water sources. Their contribution to the total water flux and chloride load varies as a function of time. In the largest part of the polder, the chloride concentration of the seepage water (diffuse seepage) is much lower than the chloride concentration of the polder water (Table 2), which implies there must be a large contribution from a saline source with higher chloride concentration.

In a simplified water and chloride balance, we demonstrate that the chloride loads observed in the Noordplaspolder can only be explained by assigning a large contribution to boil seepage. Water and chloride enter the polder via precipitation (P), diffuse seepage (q_d), paleochannel seepage (q_p), boil seepage (q_b), and the inlet of external surface water (q_i). Water leaves via pumping (q_{pump}) and evapotranspiration (ET) while chloride leaves only via pumping. The water balance is expressed as:

$$P - ET + q_i + q_d + q_p + q_b = q_{pump} \quad (\text{in M m}^3/\text{y}) \quad (1)$$

Chloride loads are calculated by multiplying the water flux by the corresponding chloride concentration c . The chloride balance is then expressed as:

$$P \cdot c_p + q_i \cdot c_i + q_d \cdot c_d + q_p \cdot c_p + q_b \cdot c_b = q_{pump} \cdot c_{pump} \quad (\text{in M kg/y}) \quad (2)$$

Filling in the values we measured (see Section 3) of P (41.2 M m³/y), ET (21.7 M m³/y), q_{pump} (33.8 M m³/y), $q_{pump} \cdot c_{pump}$ (11.6 M kg/y) and the chloride concentration c of the different water balance terms (see Table 2) into Eqs. (1) and (2) leads to the following:

$$q_i + q_d + q_p + q_b = 14.3 \quad (\text{M m}^3/\text{y}) \quad (3)$$

$$0.5 + q_i \cdot 0.223 + q_d \cdot 0.117 + q_p \cdot 0.592 + q_b \cdot 1.10 = 11.6 \quad (\text{M kg/y}) \quad (4)$$

To solve these two equations with their four unknowns, we varied water balance terms q_i , q_d , q_p and q_b randomly in all possible combinations, in such a way that their sum was equal to 14.3 M m³/y, which meant that the water balance Eq. (3) was correct. The four water balance terms were varied in the range 0–14.3 M m³/y, because their water flux for the whole polder was not known exactly and to explore all possible solutions. For each combination for which Eq. (3) was correct, the total chloride

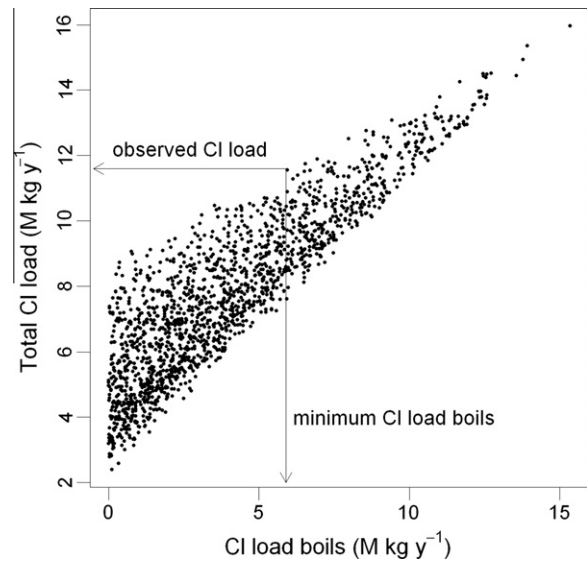


Fig. 11. Scatter diagram of calculated chloride load via boils versus calculated total chloride load in the Noordplaspolder.

load was calculated with the left-hand side of Eq. (4). Our results are shown in Fig. 11, where the calculated chloride load via boils is plotted against the calculated total chloride load. The points in Fig. 11 that show a total chloride load of 11.6 M kg/y represent the correct solutions to Eqs. (3) and (4). From the graph we can derive that the observed total chloride load of 11.6 M kg/y can only be explained with a minimum chloride load via boils of 5.8 M m³/y which is 50% of the total chloride load in the polder.

In earlier studies (ICW, 1976; Wit, 1974; Pomper and Wesseling, 1978; Griffioen et al., 2002), the seepage chloride load was obtained by multiplying the estimated seepage flux by the average chloride concentrations of groundwater at the top of the upper aquifer (–10 to –15 m MSL). This resulted in large underestimations of the total chloride load, and the discrepancy was removed by using the higher chloride concentrations of groundwater at –25 to –35 m MSL, but no justification was given. We have shown that the common assumption by the earlier studies, that upward saline seepage could be considered as a spatially uniform, diffuse process, is incorrect. We consider their choice to use higher chloride concentrations in the water and salt balances is roughly the same as our idea that more saline groundwater of deeper origin contributes to the polder discharge via boils.

5. Conclusions

From our observations and analysis of the geological structure of the Holocene confining layer, head differences, groundwater composition, surface water salinity, polder water discharge and chloride load, temperature profiles, and boil observations, in a typical deep Dutch polder (Noordplaspolder), we conclude that saline seepage is not a uniform process. We distinguished three types of seepage: (1) diffuse seepage through the Holocene layer, (2) seepage through paleochannel belts that cut through the Holocene layer, and (3) intense seepage via localized boils. These differ with regard to flux, chloride concentration, and the location in the polder where they occur. Their contribution to surface water salinization therefore also differs. The magnitude of the seepage flux reflects the depth from which the groundwater originates, with higher fluxes resulting in upconing of deeper, more saline groundwater (since the salinity of groundwater increases with depth in the Noordplaspolder).

The largest seepage fluxes and highest chloride concentrations were found in boils; these are small, localized vents in the Holocene confining layer through which groundwater discharges at high velocities. Boils were observed in ditches, near the edge of the polder, or in areas with paleochannel belts. Temperature and chloride concentration profiles demonstrated high flow velocities and strong upconing of deeper and more saline groundwater at boils, producing an average chloride concentration of 1100 mg/l. Permeable, sandy, paleochannel belts cut through the less permeable, lower part of the Holocene confining layer. This results in higher seepage fluxes through the paleochannel belts and upconing of more saline groundwater than found in the diffuse seepage occurring through the Holocene layer. Therefore, the average chloride concentration of paleochannel seepage (600 mg/l) is much higher than of diffuse seepage (100 mg/l).

The annual average chloride concentration of the polder water discharge is about 400 mg/l and may reach 720 mg/l in dry periods. Given that polder water discharge is a mixture of groundwater seepage, precipitation and inlet water, and that the latter two have a diluting effect, we set out to demonstrate that boils must be responsible for the high chloride concentrations. By integrating our field data and balance calculations, we show that boils are indeed the dominant salinization mechanism and that they contribute more than 50% of the total chloride load to the Noordplaspolder surface water.

In attempts to abate surface water salinization in deep polders, it is worth focusing attention on boils because of their dominant role in the chemical load and their local nature. Both of these features will facilitate the countermeasures needed. Since groundwater discharged via boils has a constant temperature of 11 °C, airborne thermal infrared mapping can help to localize them. Although the geohydrological situation of the Dutch polders is quite specific, our work may have broad implications for other geographical regions. Boils occur where there are significant head gradients, such as areas with e.g. dikes (Li et al., 1996; Holzer and Clark, 1993) and dammed reservoirs, but also in brook valleys or river plains. Since boils affect the surface water quality, our approach for quantifying their impact may be adopted if adequate tracers (salt or other chemical tracers, temperature) can be identified to distinguish between the different sources of water.

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