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Origin and dynamics of groundwater salinity in the alluvial plains of western Delhi and adjacent territories of Haryana State, India

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Abstract:

Groundwater salinity is a widespread problem and a challenge to water resources management. It is an increasing concern in the alluvial plains of Delhi and neighbouring Haryana state as well as a risk for agricultural production water supply and sustainable development. This study aims to identify potential sources of dissolved salts and the driving mechanisms of salinity ingress in the shallow aquifer. It combines a comprehensive review of environmental conditions and the analysis of groundwater samples from 25 sampling points. Major ions are analysed to describe the composition and distribution of saline groundwater and dissolution/precipitation dynamics. Density stratification and local upconing of saline waters were identified by multilevel monitoring and temperature logging. Bromide–chloride ratios hold information on the formation of saline waters, and nitrate is used as an indicator for anthropogenic influences. In addition, stable isotope analysis helps to identify evaporation and to better understand recharge processes and mixing dynamics in the study region. The results lead to the conclusion that surface water and groundwater influx into the poorly drained semiarid basin naturally results in the accumulation of salts in soil, sediments and groundwater. Human-induced changes of environmental conditions, especially the implementation of traditional canal and modern groundwater irrigation, have augmented evapotranspiration and led to waterlogging in large areas. In addition, water-level fluctuations and perturbation of the natural hydraulic equilibrium favour the mobilisation of salts from salt stores in the unsaturated zone and deeper aquifer sections. The holistic approach of this study demonstrates the importance of various salinity mechanisms and provides new insights into the interference of natural and anthropogenic influences. Copyright © 2011 John Wiley & Sons, Ltd.

KEY WORDS groundwater; inland salinity; semiarid; India; hydrogeochemistry; stable isotopes

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INTRODUCTION

Groundwater salinity ingress is an increasing concern in many regions worldwide and affects nature, individuals and society in many ways. It may endanger ecosystems, degrade the productivity of agricultural land, jeopardise the health and livelihood of individuals and increase the costs of water supply infrastructure and industrial processes (Villholth and Sharma, 2006; Weert *et al*., 2009). All these consequences are observed in the densely populated alluvial plains of northern India, where large regions have to deal with brackish or saline groundwater. Most affected are the semiarid to arid northwestern states of Rajasthan, Punjab, Haryana and Delhi, where groundwater consumption has been increasing over the past decades, along with population growth, agricultural revolution and industrialisation (Daga, 2003; Ambast *et al*., 2006; Villholth and Sharma, 2006). Although salinity in groundwater rises, further abstraction leads to a dramatic water-level decline in the overexploited aquifers (Rodell *et al*., 2009). To avoid further degradation of water resources and to establish sustainable management strategies, it is important to accurately assess factors of influence within a catchment and focus on both quantitative and qualitative aspects (Bouwer, 2000; Villholth and Sharma, 2006).

The aim of this study is to identify sources and processes that contribute to groundwater salinity in the study region. Initially, a comprehensive review summarises the common causes of inland salinity ingress in shallow aquifers. Field data are analysed, including temperature logs, hydrogeochemical data and stable isotope analyses from long-term studies at a field site in western Delhi and a groundwater mapping campaign in adjacent territories of Haryana state. The interpretation of local data is based on a comprehensive assessment of geographical conditions in the study region and the integration of explanatory models developed in other semiarid catchments. The article demonstrates the role of natural factors such as geomorphology, drainage and climate for the salt/water balance and highlights the effect of the man-made perturbation of the natural equilibrium through land use and water resource management. The holistic point of view demonstrates the importance of various salinity mechanisms and provides new insights into the interference of natural and anthropogenic influences.

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REVIEW ON COMMON SOURCES OF INLAND SALINITY IN SHALLOW AQUIFERS

The reasons for inland groundwater salinisation can be manifold, and several causes may contribute simultaneously (e.g. Grube *et al.*, 1999; Smith and Compton, 2004; Duncan *et al.*, 2005). Soluble ions may originate from different sources. In marine environments, salt can precipitate from seawater (marine evaporates), be trapped in the pores of sedimentary rocks (connate water) or intrude into coastal freshwater aquifers (Weert *et al.*, 2009). These salts can be stored in geological formations over large periods and present potential sources of groundwater salinity. In inland environments, dissolved ions may originate from the leaching of evaporates or be released to groundwater during mineral weathering. They can also be imported into a basin, for example, by the deposition of marine aerosols via rainfall or by the influx of mineralised surface water or groundwater (Salama *et al.*, 1999; Jolly *et al.*, 2008). Salinity ingress in shallow inland aquifers is usually a consequence of the concentration of dissolved salts through evaporation, the leaching of salts or the mobilisation of saline groundwater from deeper aquifer sections. Common processes are described as follows, and examples are illustrated schematic sketches in Figure 1:

1. Accumulation of salts. Evaporation, especially in semiarid and arid regions can lead to salinity ingress in soil and groundwater. In undrained or poorly drained basins, salts can become concentrated on surface sediments (e.g. in salt pans or playa lakes; Figure 1a) or in the soil profile (e.g. Lowenstein and Hardie, 1985; Salama *et al.*, 1999; Misra and Mishra, 2007). Percolation of residual brines or leachate leads to salinisation of shallow groundwater (Datta *et al.*, 1996; Gilfedder *et al.*, 2000). The same occurs in soils that are permanently saturated by subsurface water (waterlogged), so that evaporation of subsurface water takes place through capillary rise (Figure 1b). This is naturally observed in low-lying areas with shallow groundwater tables (Salama *et al.*, 1999; Jolly *et al.*, 2008). It can also be a consequence of anthropogenically induced groundwater-level elevations, for example, along canals or under irrigated fields (Tyagi, 1988; Seiler *et al.*, 2001; Duncan *et al.*, 2005; Singh, 2005). Excessive irrigation in semiarid and arid regions also activates processes that cause the soils to lose structure and fertility and become more susceptible to waterlogging (Salama *et al.*, 1999; Datta and de Jong, 2002).
2. Dissolution of salts. The formation of saline waters in shallow aquifers is often a result of salt leaching where evaporite-bearing soil or rock comes into contact with groundwater circulation (e.g. Cooper, 2002; Herczeg *et al.*, 1993; Smith and Compton, 2004). It may occur when these units are uplifted by structural processes (Figure 1c) including salt tectonics (Grube *et al.*, 1999; Kloppmann *et al.*, 2001). Soil or surface salts (e.g. capillary salts) can dissolve and percolate to groundwater during the infiltration of rain or irrigation water (Duncan *et al.*, 2005; Selle *et al.*, 2010). Salt stores in the unsaturated zone can also be leached after groundwater table elevation (Figure 1d), for instance, because of irrigation projects or vegetation clearance (Bradd *et al.*, 1997; Salama *et al.*, 1999).
3. Mixing with saline waters. Deep saline groundwater may have been formed by the processes described earlier. They can also be of marine origin or result from

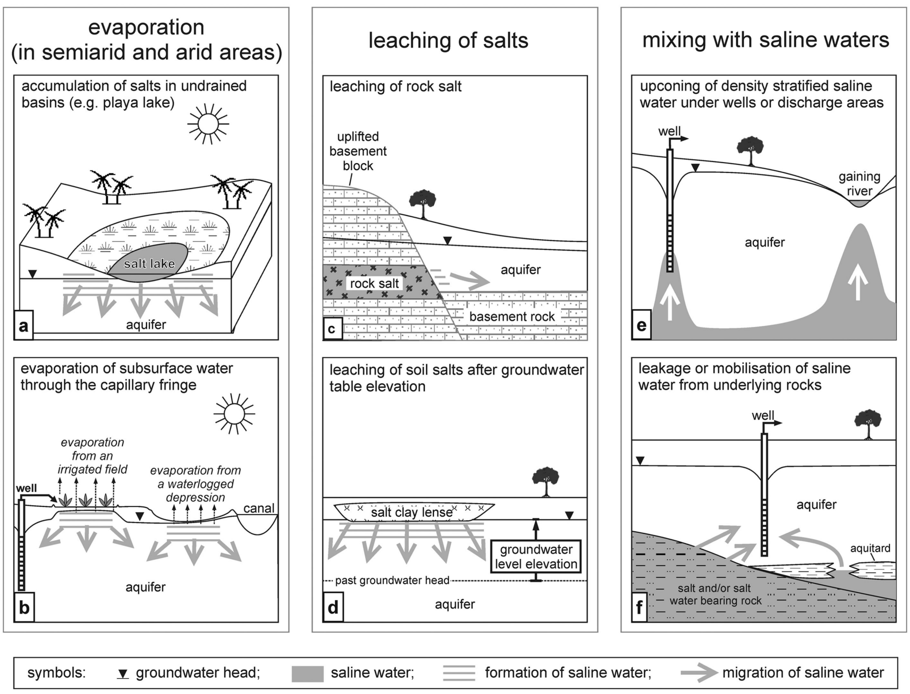

Figure 1. Schematic sketches of common sources of geogenetic inland salinity in shallow aquifers and processes leading to salinity ingress. The origin and the dynamic of saline and brackish groundwater are often controlled by a combination of these processes. They are in many cases accelerated or triggered by human-induced modifications that affect the natural equilibrium in the water cycle.

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salt concentration through membrane filtering effects at argillaceous layers (Bredehoeft *et al*., 1963) or expulsion of fluids from compacting beds (Wilson *et al*., 2003). Brines are found in deep aquifer systems of sedimentary basins all over the world and are often separated from the meteoric water cycle by impermeable layers, hydraulic traps or stable density stratification. These waters can naturally exfiltrate in a discharge area (Banner *et al*., 1989) or be mobilised and contaminate overlying freshwater aquifers through leakage and mixing (Figure 1e) (e.g. Grube *et al*., 1999). In density-stratified aquifers, a low pressure anomaly under natural discharge areas or abstraction wells can lead to upconing of the freshwater–saline water interface (Figure 1f) (e.g. Herzberg, 1901; Grube *et al*., 1999; Rao *et al*., 2007).

In semiarid and arid regions, the accumulation of salts in soil profiles or groundwater can be considerable in geologic time scales (dryland salinity). It is largely controlled by climate, morphology, surface and subsurface drainage (Bradd *et al*., 1997; Salama *et al*., 1999; Jolly *et al*., 2008). Herczeg *et al*. (1993) suggested that salts can accumulate until a quasi-steady-state equilibrium is reached where groundwater and soil water act as intermediate reservoirs. The amount of salt that is stored or washed out of these reservoirs varies appreciably in response to climatic and geomorphologic fluctuations.

THE STUDY AREA

The study area is situated in northern India and includes rural parts in the west of the National Capital Territory of Delhi and the neighbouring Haryana state (Figure 2). Field studies were mainly carried out in an area of approximately 25×15 km along the Najafgarh drain, which is a tributary of the Yamuna River.

Figure 2 illustrates that the study area situated in the alluvial plains between the rivers Sutlej (Indus catchment) and Yamuna (Ganges catchment). It is part of a geomorphologic unit, embedded between the Himalayan foreland hills in the north and the Aravalli hard rock range of the Indian Craton in the south. The low-lying terrain forms an elongated basin, with a central trench, in the broad transition zone between the catchments of the Ganga Basin in the East and the Indus Basin in the west. The Thar Desert of Rajasthan state is located southwest of the study area. Pant (1993) suggested that the desert expanded up to the Delhi region during the last glaciation, when aridity and strong southwesterly summer winds prevailed. Relict dunes in the central trench were deposited by wind from the Thar region (Subramanian and Saxena, 1983; Thussu, 2006).

The plains in the study region are mainly built up of Quaternary alluvial and locally aeolian deposits. They belong to the large aquifer system between the Himalayan Foreland and the Aravalli range. In the study area, the sediments reach a thickness of several hundred metres and consist of sands with considerable contents of silt, clay and kankar nodules (carbonate concretions). They are underlain by metamorphic hard rock of the Indian Craton (quarzites and schists with pegmatites and quartz veins) (Das *et al*., 1988; CGWB-GOI, 2006; Thussu, 2006).

The climate in the study region is of continental type and semiarid. It is characterised by a pronounced summer monsoon season from July to September, which generates more than 75% of the total annual precipitation (Thussu, 2006; CGWB-GOI, 2006). During this period, monthly mean temperatures typically lie at approximately 30 °C in Delhi. From then on, they gradually decrease until January, when monthly mean temperatures fall below 15 °C. The highest temperatures occur in the hot and dry pre-monsoon season (March–June), with monthly mean temperatures up to 34 °C (IAEA/WMO, 2009). In Delhi, normal annual rainfall (1980–2006) decreases from the central parts (710 mm with 36 rainy days per year) towards the western margin (400 mm with 19 rainy days per year) (CGWB-GOI, 2006). Climate in the study area (Figure 2) is even warmer and drier because it lies in the transition zone towards the Thar Desert (Pant, 1993, Tripathi and Rajamani, 1999, Thussu, 2004).

Owing to the semiarid conditions, groundwater recharge from rain is only significant during monsoon seasons. A major share of infiltrating rainwater and flood water is retained on the land surface or in the unsaturated zone (Das *et al*., 1988, Datta *et al*., 1996, Kumar *et al*., 2008).

Estimates of fractional recharge from rain to groundwater range from approximately 15% in semiarid Haryana (Goel *et al*., 1977) to less than 5% in most parts of Delhi (Datta and Tyagi, 2004). A major contribution to aquifer recharge originates from canal/river seepage and infiltration of agricultural and urban surface runoff that accumulates in stagnant water pools and seeps slowly into the groundwater (Datta *et al*., 1996). Groundwater discharge towards the Indus and Ganga basins is constricted owing to low surface gradients and limited permeability of the alluvial aquifer (Thussu, 2004).

Surface water discharge from the study region is almost negligible, except after monsoon rains. Figure 2 visualises that morphology in the Yamuna-Sutlej interfluve does not allow the watershed between the Ganges and the Indus catchments to be clearly determined. In the central trench, local depressions (topographic height less than 215 m) represent small basins, some without surface drainage. Many small streams abort with dead ends in their terminal fan deposits (Srtvastava *et al*., 2006; Thussu, 2006). The Najafgarh Drain, which flows through the study area, is subdivided by a weir at the interstate border. The upper segment in Haryana State falls dry in summer but receives surface runoff and even storm water from the Sahibi catchment in the Aravalli range during monsoon (AGC, 2001). The lower segment in Delhi was canalised in 1978 to drain a morphological depression in western Delhi (Najafgarh Jheel) and was integrated into the storm water and sewer drainage system of the city (Nema and Agarwal, 2003).

Despite the semiarid conditions, the fertile soils of the densely populated plains are intensively used for agriculture. This is only possible with extensive irrigation (Ambast *et al*., 2006). Since ancient times, the natural drainage system has largely been modified for irrigation and interbasin water

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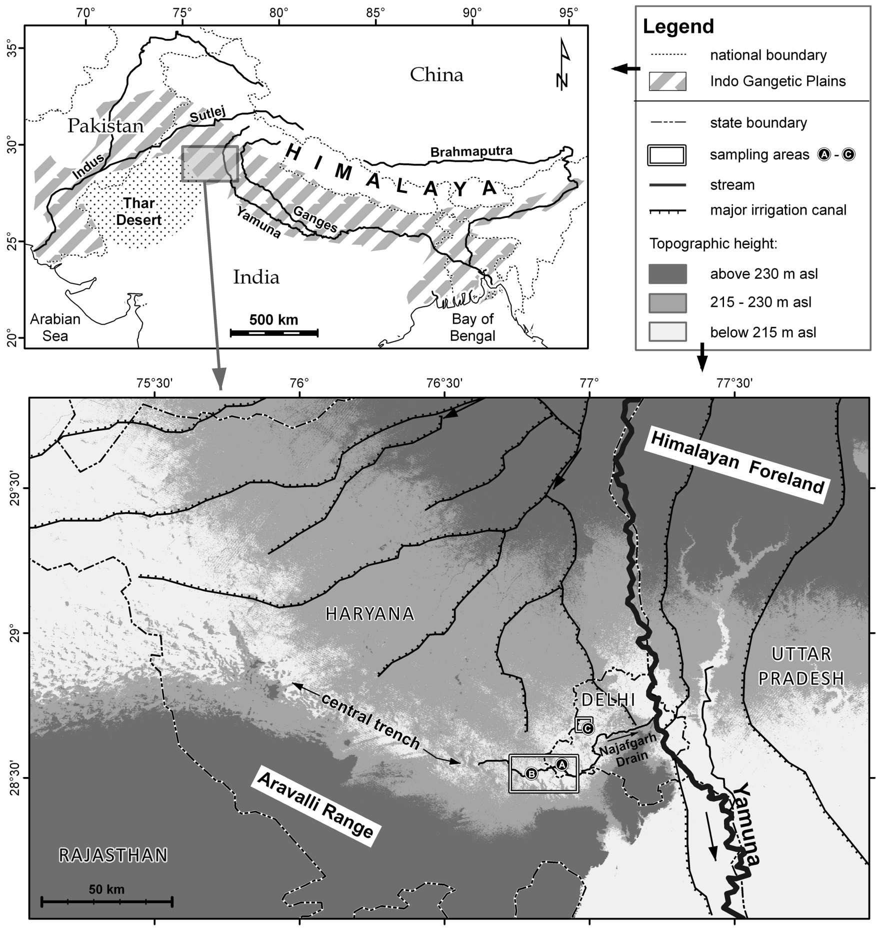

Figure 2. Location of the study areas (sampling areas A–C) and topographic features of the study region. The low-lying terrain is embedded between the Himalayan range in the north and the denudated Aravalli range in the south. The elongated depression forms part of the Indo-Gangetic alluvial plains and is situated in the broad intersection zone of the catchments of both streams. A network of irrigation canals distributes Himalayan waters across the semiarid terrain.

transfer (Singh, 2005). A network of irrigation canals distributes Himalayan water all over the semiarid terrain. Most of them end up as dead-end canals in the plains (Figure 2). Since the 'green revolution' in Indian agriculture, traditional canal irrigation has been widely replaced by more productive groundwater irrigation (Ambast *et al*., 2006). Over the past decades, population growth and changes to more profitable crops have boosted water demand. Today, the plains of Delhi and Haryana have a high performance in terms of agricultural productivity. The system is anything but sustainable, however, because groundwater levels are declining (Rodell *et al*., 2009) and salinity ingress is a widespread problem (Datta *et al*., 1996; Villholth and Sharma, 2006; Singh, 2005; Misra and Mishra, 2007).

METHODS

This study combines field data collected during two different field studies (Table I). The Delhi site (area A) was built up for

Table I. Sampling areas and methods

Sampling areas	Delhi field site (A)	Haryana mapping area (B) and Delhi sampling point no. 25 (C)
Primary purpose of investigation	Monitoring groundwater–surface water interaction	Groundwater sampling and salinity mapping
Sampling points	Multilevel monitoring wells (~15 m north of the drain) and local dug well (~500 m north of the drain)	Local wells (e.g. irrigation wells, traditional dug wells, hand pumps), unsystematically built
Sampling depth below ground level	Multilevel: 3.5–9.5 m (shallow), 18–24 m(medium) and 31–37 m (deep); dug well: 0–5.3 m	Variable depths up to ~40 m (details unknown); often with large filter screens
Sampling technique	Submersible 12-V pump	Local equipment (e.g. diesel engine suction pumps, bucket, pumping rod)
Sampling period	Repeated campaigns from December 2006 to February 2008	One single campaign March 2008 to April 2008

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long-term monitoring of groundwater–surface water interaction, described by Lorenzen *et al.* (2010). In the Haryana area (area B), local wells were sampled with the intention of identifying the spatial distribution and composition of saline groundwater. An additional sample (sample 25, area C) was taken some 20 km northeast at Tiki Kalan village, which is a hot spot in terms of groundwater salinity (Datta *et al.*, 1996; Kumar *et al.*, 2008).

During groundwater sampling, physicochemical parameters (pH, redox potential, electric conductivity, temperature and dissolved oxygen concentration) were monitored on site, and alkalinity was determined with titration field kits (Aquamerck). Samples for the analysis of anions (filtered with cellulose acetate membrane, 0.2 μm), cations (filtered and acidified with nitric acid) and stable isotopes were stored separately in 20 ml of polyethylene bottles. At the Geosciences laboratories of Freie Universität, inorganic ions (of Ca, Mg, Na, Cl, SO_4, NO_3 and Br) were analysed with a photometer (Technicon Autoanalyser), an ion chromatograph (Dionex DX 500) and an inductively coupled plasma mass spectrometer (ICP Perkin Elmer Otima 2100). Ion activities and carbonate species were calculated with PHREEQC software (Parkhurst, 1995). Stable isotope ratios (δ^2H, $\delta^{18}O$) were determined with a Thermo Finnigan Mat 253 isotope ratio mass spectrometer. Temperature–depth profiles at the Delhi field site were recorded with a data logger unit (Solinst Levellogger) with a high-resolution temperature and pressure conducer that was gradually lowered into the well.

RESULTS AND INTERPRETATION OF FIELD DATA

Spatial distribution of groundwater salinity

The locations of the sampling points (Table II) and salinity ranges of the corresponding groundwater samples are shown in Figure 3. According to their totally dissolved solids (TDS), the samples are classified as proposed by Davis and DeWiest (1966) as fresh (<1000 mg/l), brackish (1,000–10,000 mg/l) or saline (>10,000 mg/l). Brackish waters are further subdivided

Table II. Groundwater samples and the results of hydrogeochemical and stable isotope analysis

No.	Site	Type	Date	Na^+	Ca^+	Mg^+	Cl^-	SO_4^{2-}	HCO_3^-	Br^-	NO_3^-	$\delta^{18}O$	δ^2H	TDS
(A) Delhi field site														
1a	Shallow I	OW	12/06	560	170	61	890	290	506	–	0.1	−1.4	−22	2,500
1b	Shallow I	OW	05/07	540	180	53	750	270	510	–	0.1	−1.2	−20	2,200
1c	Shallow I	OW	02/08	880	260	100	1,400	320	–	2	4	−1.9	−25	2,900
2a	Medium	OW	12/06	2100	500	260	3,800	1300	320	–	0.5	−4.0	−33	8,200
2b	Medium	OW	05/07	2200	540	280	3,900	1300	320	8	4	−3.8	−31	8,600
2c	Medium	OW	02/08	2600	630	330	4,700	1400	–	5.9	24	−3.8	−32	9,700
3a	Deep	OW	12/06	3200	680	440	6,600	1000	260	–	0.5	−4.7	−38	13,000
3b	Deep	OW	05/07	3400	740	460	6,700	1400	240	9	3	−4.5	−38	13,000
3c	Deep	OW	02/08	4800	870	550	9,300	1500	–	8.5	2	−5.0	−38	17,000
4a	Dug well	DW	12/06	800	190	190	1,500	400	730	–	22	−3.2	−29	3,900
4b	Dug well	DW	05/07	630	150	130	1,100	270	710	–	6	−3.4	−31	3,000
4c	Dug well	DW	02/08	–	–	–	1,500	470	–	2.3	14	−3.1	−30	–
(B) Haryana mapping area														
5	Ismailpur	TW	04/08	54	69	43	87	8	340	0.4	37	−6.2	−50	650
6	Bridge 8-1	TW	04/08	280	150	62	520	180	-	1.6	2	-	-	1,200
7	Bamlad	TW	04/08	5400	890	1300	11,000	4300	230	21	15	−5.0	−46	23,000
8	Budhera	TW	04/08	27	35	24	6	5	210	0.2	21	−5.9	−53	350
9	Centre	HP	04/08	150	62	43	45	17	700	0.3	1	−4.3	−40	1,000
10	DAD FW	TW	04/08	430	78	67	370	270	510	1.6	67	−4.9	−37	1,800
11	Daria	TW	04/08	1200	170	360	2,100	1200	400	5	45	−4.9	−36	5,500
12	HP Drain	HP	04/08	230	78	91	470	100	480	1.4	0.7	−5.8	−44	1,500
13	Kurk	TW	04/08	760	390	370	2,600	760	350	6	71	−4.0	−37	5,400
14	LAG-DE	TW	04/08	690	340	230	1,800	380	350	4	17	−3.1	−30	3,800
15	Mubarik	HP	04/08	200	170	160	580	140	360	2.8	150	−5.4	−41	1,800
16	Munur	TW	04/08	150	32	52	90	36	490	0.5	23	−4.9	−41	890
17	Pelpa	HP	04/08	680	100	160	1,000	280	780	3.7	72	−5.6	−47	3,100
18	Sondhi	TW	04/08	1200	260	390	3,100	610	740	12	91	−5.5	−37	6,100
19	Sultanpur1	TW	04/08	150	66	55	270	31	330	0.6	60	−6.0	−47	980
20	Sultanpur2	HP	04/08	310	33	35	98	190	590	0.4	50	−5.6	−42	1,300
21	Weir-FW	TW	04/08	360	200	160	1300	110	340	3	6	−5.3	−39	2,500
22	Weir-HP	HP	04/08	220	59	55	280	34	513	0.7	20	−5.5	−45	1,200
23	Workers	TW	04/08	310	26	44	200	190	-	0.6	0.1	-	-	800
24	Dadri	HP	04/08	250	67	55	320	190	340	2	35	-	-	1,300
(C) Additional sample														
25	Tiki Kalan	TW	04/08	1900	850	730	5500	2000	240	6	9	−5.6	−43	11,000

Units: concentrations in parts per million; stable isotope ratios in ‰ (vs V-SMOW).
Sample type: OW, observation well; TW, private tube well (motor pump); HP, hand pump; DW, traditional open dug well.

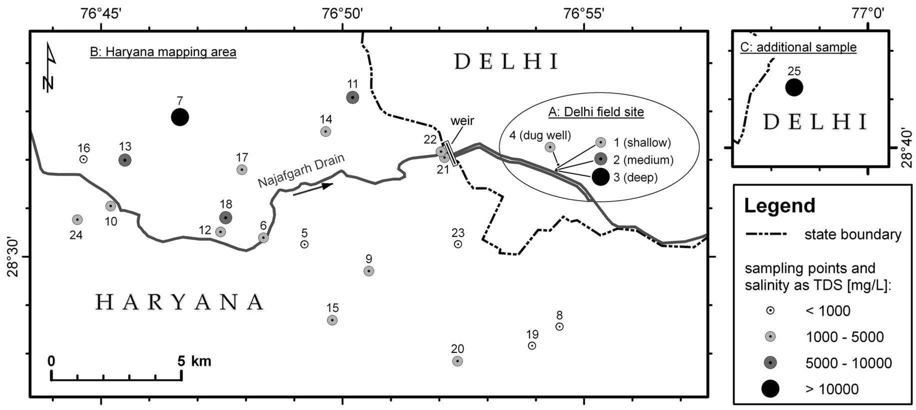

Figure 3. Location of groundwater sampling points in the study areas and salinity ranges (for regional context compare Figure 2).

into weakly brackish (TDS from 1000 to 5000 mg/l) and highly brackish (TDS from 5,000 to 10,000 mg/l). Mapping the TDS values illustrates that groundwater salinity is generally higher in the north of the Najafgarh drain. In the southern part, between the drain and the hard rock ridge, most samples are weakly brackish or fresh.

With regard to the samples from the Haryana area, it has to be considered that sampling depths are mostly unknown (Table I) and large filter screens may lead to the mixing of groundwater from different depths. The importance of the well-designed information becomes clear from the results of the Delhi field site, where multidepth sampling was possible and deeper groundwater is considerably more saline.

Temperature profiles and depth-dependent TDS concentrations from the Delhi field site are shown in Figure 4. Temperature oscillations in the upper 5 m below groundwater head reflect seasonal variations of the atmosphere. This surficial zone is relatively narrow, as is typically the case in discharge areas (Anderson, 2005). Below the surficial zone, the temperature increases by 0.2 °C/ m from around 28.5 °C at 10 m b.g.l. to 34 °C at 37.5 m b.g.l. The gradient of this geothermal zone is very high compared with gradients measured elsewhere in Delhi (Lorenzen *et al.*, 2010). Between 25 and 37 m b.g.l., the curve is not a straight line but a convex upward curve. Both a steep gradient and a convex upward profile indicate upward flow of relatively warm water in discharge areas (Anderson, 2005).

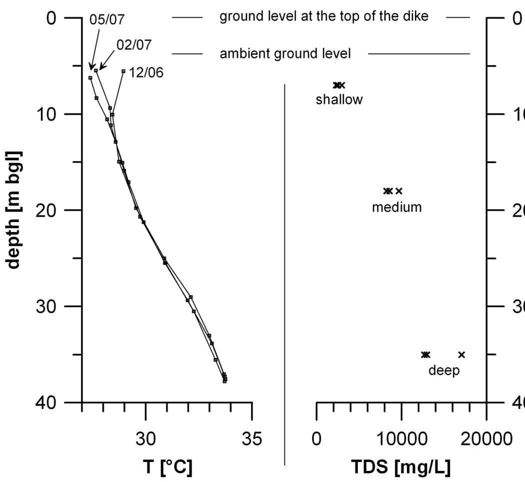

Figure 4. Temperature profiles and depth-dependent salinity in groundwater at the Delhi field site.

Salinity increases with depth, from slightly brackish in the shallow well to highly brackish in the medium well and saline in the deep one. In the medium and deep groundwater, salinity significantly increases during the three sampling campaigns, probably because of the upconing of saline groundwater.

Major ions and salinity

The most abundant solutes in groundwater are the ions of sodium NaCl, CaSO4 and $CaCO_3$. The anions of these major salts are plotted against salinity (as TDS) in Figure 5. The samples range from fresh to saline, but most of them are classified as weakly brackish with TDS between 1000 and 5000 mg/l. Salinity clearly correlates with Cl^- and SO_4^{2-} in a linear regression with the determination coefficients $r^2 = 0.98$ and $r^2 = 0.95$, respectively. HCO_3^- does not increase with salinity and remains in a range between 3 and 11 mEq/l in all samples. It is the dominant anion in the freshwater, whereas Cl^- becomes the most frequent anion in brackish and saline water. SO_4^{2-} increases along with Cl^-, but the equivalent concentrations are in general several times lower and more scattered. It is interesting to note that SO_4^{2-} concentrations remain below HCO_3^- in most of the brackish waters with TDS less than 5000 mg/l. Above this limit, concentrations of HCO_3^- decline whereas those of SO_4^{2-} increase.

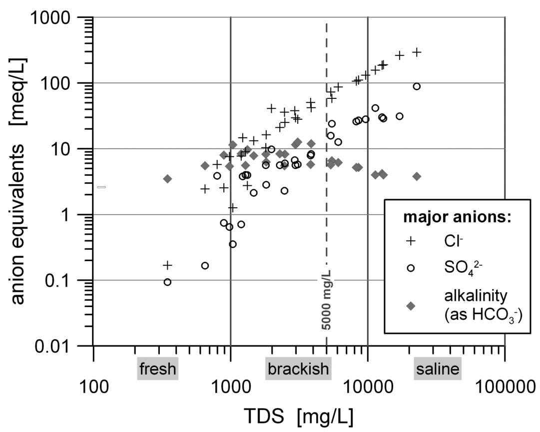

Figure 5. Salinity (as TDS) and concentrations of the anions of major salts.

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Diagrams in Figure 6 show relationships between the dissolved ions of $CaSO_4$ and $CaCO_3$. The dissolution–precipitation equilibriums of these salts affect each other substantially because both salts compete for the same cation.

Primary and secondary calcite minerals (e.g. detrital carbonate grains, kankar concretions, grain coatings) are abundant in the alluvial aquifer (Courty and Fédoroff, 1985, Yadav and Rajamani, 2004; Thussu, 2006), so groundwater should in general be saturated with respect to $CaCO_3$. On the logarithmic plot in Figure 6a, the equilibrium condition is a straight line. The slope depends on ionic strength (i) of the solution and partial pressure of carbon dioxide (pCO_2). Such a line was calculated according to Appelo and Postma (2005) and plotted for an exemplary brackish groundwater with $i = 0.18$ (e.g. sample 4) under CO_2 partial pressure $10^{-1.5}$ atm (typical value for soil). Under these conditions, the field above the equilibrium line represents supersaturation, and the field below the line shows subsaturation. Most samples plot close to the line or below it. Those with a high Ca^{2+} content are generally low in HCO_3^- and vice versa because $CaCO_3$ saturation controls the maximum concentrations of the ion product of $Ca^{2+} \times (HCO_3^-)^2$. The samples that plot far below represent mostly fresh groundwater from the area south of the drain. This region is recharged by waters that infiltrate further south in the Aravalli hard rock hills (Gupta, 2008; Figure 2). In the denudated hills, soil cover is scarce, so CO_2 pressure during groundwater recharge is low and thereby also the uptake of HCO_3^- and the potential to dissolve calcite.

$CaSO_4$ is much more soluble than $CaCO_3$, so saturation is not reached. Ca^{2+} and SO_4^{2-} correlate positively in the brackish and saline samples (Figure 7b). Most samples plot around the trend line with a molar ratio of $r = 1$, representing the dissolution of $CaSO_4$. It is therefore concluded that Ca^{2+} concentrations in the brackish and saline groundwater are mainly controlled by the amount of dissolved $CaSO_4$. Owing to the common ion Ca^{2+} in $CaCO_3$ and $CaSO_4$, the dissolution of $CaSO_4$ has a direct effect on $CaCO_3$ saturation. Increasing Ca^{2+} concentrations may even trigger the precipitation of $CaCO_3$ when the ion activity product of Ca^{2+} and HCO_3^- exceeds saturation (eq. 1).

$$\begin{aligned} Ca^{2+} + 2\,HCO_3^- &\rightarrow \mathit{CaCO_3} \downarrow + H^+ + HCO_3^- \\ &\rightarrow CaCO_3 \downarrow + CO_2 \uparrow + \mathit{H_2O} \qquad (1) \end{aligned}$$

The equation also shows how the precipitation of $CaCO_3$ is accompanied by the increase in H^+ in a first step and the formation of CO_2 in a second step. When gypsum solution

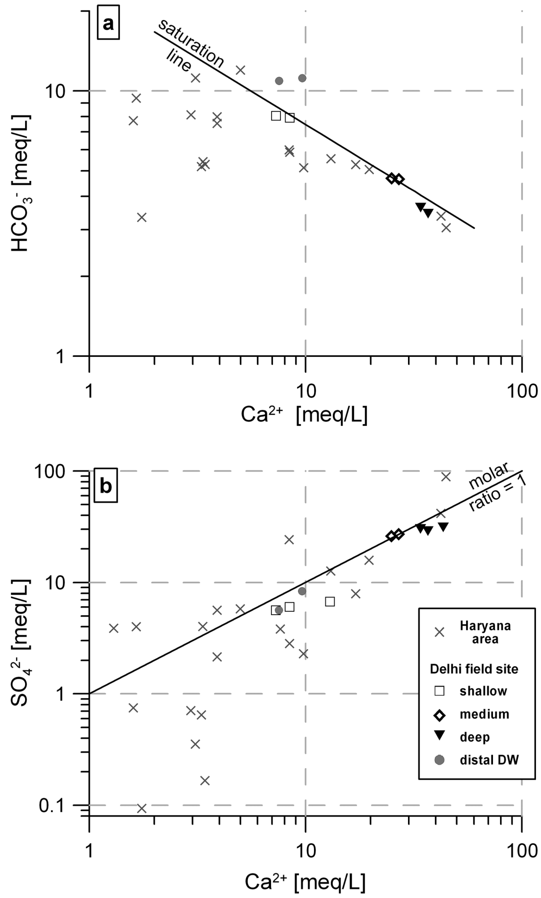

Figure 6. Ca^{2+} concentrations in relation to the anions of its major salts. Ca^{2+} concentrations in general increase with salinity, and many samples plot close to the $CaSO_4$ dissolution line (molar ratio = 1). In the HCO_3^- versus Ca^{2+} diagram, most samples plot on or below the $CaCO_3$ saturation line (calculated for an exemplary solution at $pCO_2 = 10^{-1.5}$ and ionic strength $i = 0.18$). The equilibrium with the $CaCO_3$ solid phase probably imposes a constraint on the ion product of $[Ca^{2+}] \times [HCO_3^-]^2$.

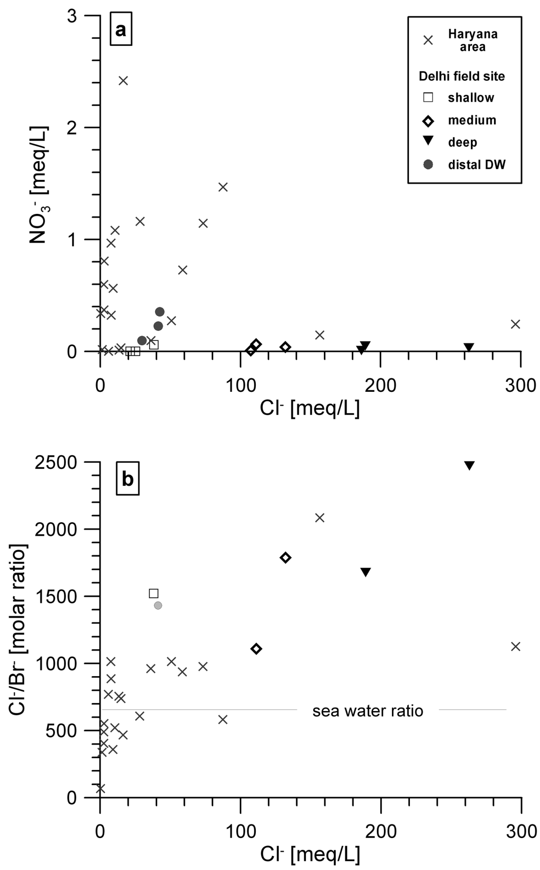

Figure 7. Correlations of Cl^- salinity and indicator ions NO_3^- and Br^-. In the rural area, NO_3^- mainly originates from fertilisers and seeps to groundwater with irrigation return flow. Cl^-/Br^- ratios are increased in the brackish and saline groundwater, suggesting that the dissolution of salts is a significant contributing factor to groundwater salinity.

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and calcite precipitation take place in shallow groundwater, excess CO_2 can be lost to the atmosphere. Mixing $CaCO_3$ saturated saltwater, rich in $CaSO_4$, would cause the same reaction.

In deeper aquifer sections, increased hydraulic head permits the dissolution of additional CO_2. The release of hydrostatic pressure (i.e. through uplift during groundwater discharge), however, must lead to the degassing of CO_2. In this context, it is interesting to note that the massive formation of gas bubbles (non-odorous, non-flammable) was observed during sampling at the deep well at the Delhi field site.

Br^- and NO_3^- as indicator ions

The concentrations of certain ions can help to further constrain hydrogeochemical processes related to groundwater salinity. Although NO_3^- is often an indicator for anthropogenic influence, characteristic Cl^-/Br^- ratios allow conclusions to be drawn about the formation of salinity.

In the rural aquifers around Delhi, NO_3^- is mainly derived from fertilisers and is typically associated with irrigation return flow to groundwater (Datta *et al*., 1997; Kumar *et al*., 2008). In the groundwater samples, nitrate was the dominant N-species, but it has to be considered that it does not behave conservatively and can naturally be attenuated over time by denitrification. The NO_3^- versus Cl^- diagram (Figure 7a) shows an interesting contrast: the samples at the Delhi field site are low in NO_3^- (except for those of the distal dug well, which is surrounded by agricultural fields). Adjacent to the drain, groundwater seems to be unaffected by recent agricultural contamination with nutrients. It may be ascending from deeper aquifer sections, as concluded from Figure 4. In the samples from the Haryana area, by contrast, NO_3^- is often increased, with contents up to 150 mEq/l. Between some samples, there seems to be a positive correlation between Cl^- and NO_3^-, but overall high NO_3^- is not necessarily linked with high salinity. Samples with the highest Cl^- concentrations (>100 mEq/l) are by contrast relatively low in NO_3^-, and many samples with low Cl^- (<20 mEq/l) have high NO_3^- contents. This is evidence of the complexity of the interference of anthropogenic activities and natural processes. Agriculture can, for instance, increase salinity along with nutrient contents, when irrigation waters are concentrated through evaporation or when groundwater irrigation triggers the mobilisation of brackish/saline groundwater from deeper aquifer sections (Datta *et al*., 1997; Kumar *et al*., 2008). On the other hand, freshwater is preferred for irrigation and therefore more likely to be contaminated with NO_3^-.

Chloride–bromide ratios (Cl^-/Br^-) as plotted in Figure 7b are a conventional method to analyse the origin of saline water. Both anions usually behave conservatively in groundwater (Davis *et al*., 1998). The Cl^-/Br^- ratio in precipitation is usually below the value of seawater (molar quotient ~ 650) and is often reflected by similar ratios in shallow groundwater (Davis *et al*., 1998). When continuous evaporation leads to the precipitation of NaCl, Br^- preferably remains in the residual brine. High Cl^-/Br^- ratios (molar ratios of several thousand) in groundwater therefore typically indicate the dissolution of rock salt (Herczeg *et al*., 1993; Alcala and Custodio, 2008).

In Figure 7b, many samples are depleted in bromine and plot above the seawater line, especially those with highest Cl^- salinity. However, Cl^-/Br^- ratios less than 2500 in all samples and even less than 1000 in most of them are too low to represent the dissolution of massive rock salts as shown in Figure 1c. Possibly, they have formed through the dissolution of salt deposits that contain evaporate salt but to some extent also residues of the non-evaporated fraction. Such deposits could either exist in the unsaturated zone (Figure 1d) or in sediments in deeper aquifer sections (Figures 1a and 1f). Leaching evaporate minerals from dunes of the adjacent Thar Desert, as suggested by Subramanian and Saxena (1983), would also increase Cl^-/Br^- ratios.

The less saline samples ($Cl^- < 100$ mEq/l) have Cl^-/Br^- ratios around the seawater ratio line with considerable scattering. The anion quotient may have been conserved from rainwater. Many natural processes in the soil (e.g. evaporation, transpiration, mixtures of waters with similar ratios) change the absolute concentrations but should not significantly modify the Cl^-/Br^- ratio of not-too-saline groundwater (Alcala and Custodio, 2008). The depletion of Br^- (Cl^-/Br^- ratios >650) can result from the dissolution of soil salts or mixing with more brackish or saline water. Trends are not very clear, possibly because natural Br^- concentrations can be altered by different processes. This could be, for instance, the uptake/release of Br^- by soil organic matter, anion adsorption and exchange or pollution through some anthropogenic contaminants, for example, Br^- based pesticides (Gerritse and George, 1988; Davis *et al*., 1998; Alcala and Custodio, 2008).

Stable isotopes

Oxygen and hydrogen isotope ratios ($\delta^{18}O$, δ^2H) are plotted in Figure 8a. The local meteoric water line (LMWL) is a fractionation line that represents the relationship between $\delta^{18}O$ and $s\delta^2H$ in local rainwater. It was calculated from the weighted annual mean of precipitation in New Delhi (1961–2001) from the IAEA/WMO (2009) database. All groundwater samples plot below the Delhi LMWL, with a linear relationship, in the following called groundwater line. Das *et al*. (1988) reported a similar trend from groundwater samples from different parts of Delhi. The deviation of the groundwater line from the rainwater line (LMWL) is assumed to result from non-equilibrium fractionation during evaporation. Evaporative enrichment therefore must have a major effect on the isotopic composition during groundwater recharge.

The intersection point of the LMWL and the groundwater line ($\delta^{18}O = -8.4$‰, $\delta^2H = -58$‰) indicates the medium composition of the water sources that recharged the aquifer.

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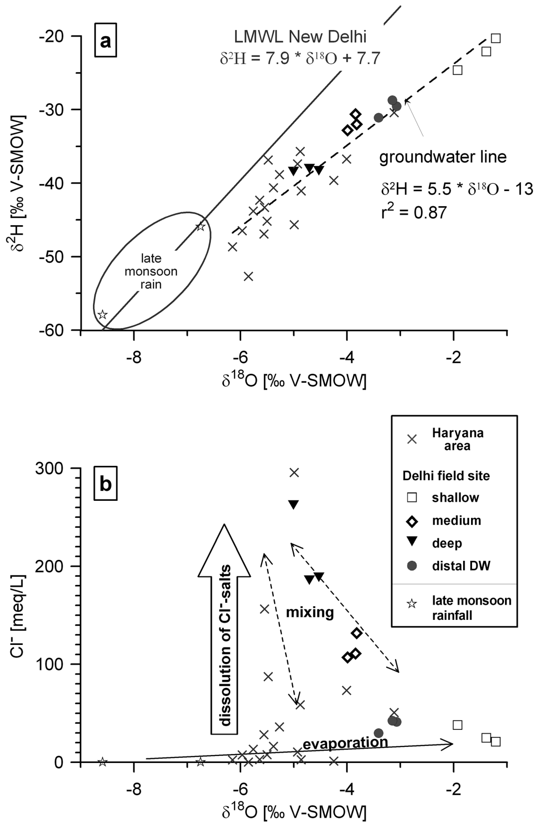

Figure 8. Stable isotope composition and groundwater salinity. In the $\delta^{18}O$ versus $\delta^{2}H$ diagram, most samples plot along the groundwater line that intersects with the LMWL in the range of late monsoon precipitation. Where enrichment in heavy isotopes correlates with Cl^-, evaporation may be a driving factor of salinity ingress. Most saline waters do not have very negative δ18O values, so their salinity is rather a result of the dissolution of salts.

This value is much more negative than the IAEA/WMO (2009) weighted annual mean precipitation in Delhi ($\delta^{18}O = -5.4$‰, $\delta^{2}H = -35$‰). It lies in the range of the IAEA/WMO (2009) data for late monsoon rainfall (August: $\delta^{18}O = -6.8$‰, $\delta^{2}H = -46$‰; September $\delta^{18}O = -8.6$‰; $\delta^{2}H = -58$‰), which is more depleted in heavy isotopes owing to an amount effect (Clark and Fritz, 1997; Datta *et al.*, 1991). Late monsoon rain could be the dominant source of aquifer recharge because monsoon rain (July–September) makes up approximately 80% of the total annual precipitation (CGWB-GOI, 2006). Rainfall in the rest of the year does not considerably contribute to aquifer recharge (Kumar *et al.*, 2008). Early monsoon rain (July) falling on parched ground is less likely to penetrate to the groundwater (Bhattacharya *et al.*, 1985; Datta *et al.*, 1996).

Another potential source of groundwater recharge is irrigation canal water, diverted from the Himalayan rivers (Datta *et al.*, 1997; Singh, 2005). Owing to the continental and altitude effect in isotope fractionation, the annual mean of these source waters is depleted compared with Delhi rainfall. In the headwaters of the Yamuna River, for instance, $\delta^{18}O$ ranges from -6‰ to -10‰ (Dalai *et al.*, 2002).

In Figure 8a, the groundwater from different sampling points plot over a wide range along the groundwater line. The samples from the Haryana area are less influenced by evaporation and scatter considerably. Groundwater at the Delhi field site is most enriched in heavy isotopes. Especially, the samples from the shallow observation wells show evidence of strong evaporation. They represent shallow groundwater in the morphological depression along the drain where evaporation may result from either water-logging where the water table is close to the soil surface or seepage from canal-irrigated fields.

In Figure 8b, $\delta^{18}O$ is plotted against Cl^- to visualise relationships between isotopic composition and salinity. The concentration of salts through evaporation should be indicated by Cl^- contents increasing along with $\delta^{18}O$ enrichment in a linear correlation (Clark and Fritz, 1997). By contrast, the dissolution of salts leads to salinity ingress but not to the fractionation of the water molecules. (The samples scatter considerably within a triangular shape, and three members can be identified:

- Samples in the left bottom of the triangle (with Cl^- concentrations less than 100 mEq/l and $\delta^{18}O$ values less than -5‰) are interpreted as fresh groundwater, recharged from monsoon rain or canal seepage.
- Samples at the top of the triangle, with similar $\delta^{18}O$ values and high chloride contents (e.g. the deep water from the Delhi field site), can be formed through the dissolution of Cl^- salt by fresh groundwater.
- Samples in the right corner are enriched in heavy isotopes and salts because of evaporation.

All samples within the triangle can be influenced by the evaporation (shifting towards the right) and/or the dissolution of salts (shifting upwards). In addition, other processes may have a significant effect, especially mixing different end members, the seasonal cycles of dissolution/precipitation in the soil and the transpiration (Datta et al, 1996; Kumar *et al.*, 2008).

Samples from the dug well and the shallow observation wells at the Delhi field site follows an evaporation trend as well as some samples from the Haryana area. By contrast, groundwater with the highest Cl^- contents (samples 3a–c, 7 and 25 in Table I) are much less influenced by evaporation; thus, their dominant source of salinity must be the dissolution of salts or mixing with a highly saline end member.

DISCUSSION OF FIELD RESULTS AND SALINITY SOURCES IN THE REGIONAL CONTEXT

The results of the water analysis have shown that there is a large variability in groundwater composition and suggest that more than one process contributes to groundwater salinity. Mixing multiple-sourced groundwater with variable salinities as concluded by Datta *et al.* (1996) for the Delhi region seems plausible. In this chapter, possible sources of groundwater salinity as shown in Figure 1 are discussed in relation to the study region. Geographic realities of the region and results from the field studies are taken into account to discuss which of these processes can be responsible for groundwater salinity and increasing mineralisation trends in the shallow aquifers.

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Accumulation of salts

In the study region, geographical conditions favour the concentration of salts. Dissolved minerals are carried into the study region by surface and groundwater flux from the Himalaya and Aravalli ranges (Figure 2). Warm and dry climate results in high potential evaporation, whereas low surface gradients and relatively fine-grained sediments limit the flushing of solutes (Datta *et al*., 1996; Thussu, 2004; Misra and Mishra, 2007).

In the groundwater samples, evaporation is reflected by the enrichment in heavy isotopes, following an evaporation line as shown in Figure 8a. Before infiltration, evaporation from open water bodies, namely, reservoirs, irrigation canals and flood-irrigated fields, amplifies the concentration of dissolved salts (Singh, 2005; Misra and Mishra, 2007; Jolly *et al*., 2008). Lateral wash off and pickup of soil salts can mobilise considerable amounts of salts in semiarid catchments (Duncan *et al*., 2005; Gilfedder *et al*., 2000). Overland flow, enriched in salts and heavy isotopes, collects in low-elevation areas and percolates to groundwater (Datta *et al*., 1996).

Salt lakes, as shown in Figure 1a, are presently not found in the study region but occur to the southwest in the Thar Desert (Yadav, 1997). They have often formed in abandoned channels or river segments that dried up because of neogene tectonics (Roy, 1999; Valdiya, 2002; Srtvastava *et al*., 2006). Palaeochannels are abundant in the study region, especially in the central trench of the study area (Singh, 2005). In similar environments, evaporite sequences may have been deposited in the study region in the geologic past, especially during periods of increased aridity, for instance during the last glaciation (Thussu, 2004). Regional climate must have been more arid whenever the monsoon weakened or failed, as for instance during the last glaciation (Pant, 1993; Andrews *et al*., 1998). Under such conditions, considerable salt stores may have built up in low permeable sediments (Lowenstein and Hardie, 1985). Kulkarni *et al*. (1989) also suggest that connate saline waters may have formed during a drier interglacial phase in the Pleistocene.

Whereas the deposition of salt stores in playa or lake bed sediments seems possible (Thussu, 2004), residual brines from salt lakes or waterlogged areas cannot be the primary source of salinity. Such concentrates would be highly enriched in heavy isotopes, as for instance at the Sambhar salt lake in Rajasthan, where $\delta^{18}O$ in subsurface brines correlates with salinity and reaches values up to +7‰ (Yadav, 1997). In the study area, by contrast, $\delta^{18}O$ in groundwater remains negative, there is no correlation with salinity and the most saline samples are only slightly enriched in heavy isotopes (Figure 8b).

Whereas salt lakes are not relevant in recent times, the evapotranspiration from waterlogged areas and irrigated fields (Figure 1b) is an increasing concern in the plains and often directly connected to anthropogenic activities (Tyagi, 1988; Datta and de Jong, 2002; Singh, 2005; Kumar *et al*., 2008). It is generally accepted that the construction of the canal network and subsequent water table elevation, waterlogging and (over)irrigation are a major source of secondary salinisation in soils and shallow groundwater in the Indo-Gangetic plains (Tyagi, 1988; Seiler *et al*., 2001; Datta and de Jong, 2002; Singh, 2005). Over the past decades, tube well irrigation has increased the temporal and spatial availability of irrigation water (Ambast *et al*., 2006; Villholth and Sharma, 2006; Gupta, 2008). In the plains of Haryana and Delhi, enormous amounts of groundwater are lifted to the surface and exposed to evapotranspiration during infiltration. Recycling the return flow to groundwater for irrigation gradually decreases the water quality of the reservoir (Tyagi, 1988; Datta *et al*., 1997; Singh, 2005; Datta and de Jong, 2002; Kumar *et al*., 2008).

Groundwater irrigation and waterlogging in the study region are expected to be two reasons for increased Cl^- and NO_3^- contents of brackish water samples. The most obvious examples are the shallow groundwater samples from the Delhi field site, which are most depleted in light isotopes, because of waterlogging in the morphologic depression. The most saline water samples ($Cl^- > 100$ mEq/l) are by contrast not much enriched in heavy isotopes and not contaminated with NO_3^-, so the high salinities cannot be a direct consequence of agricultural activities or evaporation.

Dissolution of salts

Cl^-/Br^- ratios in the groundwater samples are too low to represent the leachate of massive halite rock, and there are no reports about major marine evaporite units in the study region. The dissolution of massive rock salt, as for instance shown in Figure 1c, cannot therefore be the source of salinity. However, the existence of interlayered evaporite sequences in the sedimentary basin cannot be excluded (Thussu, 2006). Playa or salt lake deposits such as salt clay could for instance contain evaporate salt but to some extent also residues of the non-evaporated fraction (Lowenstein and Hardie, 1985; Kulkarni *et al*., 1989). Such deposits could either exist in the unsaturated zone (Singh, 2005) or in playa-like deposits (i.e. salt clay) in deeper aquifer sections (Thussu, 2004). Bromide would be depleted in the minerals but enriched in the connate brines. If both solid phase and liquid residues are stored in the same basin sediment (e.g. salt clay), flushing these sediments would not result in very high Cl^-/Br^- ratios.

Salt deposits in fine grained sediments could be preserved from rapid flushing and form saline interlayers in the sedimentary basin. Owing to their low permeability and low hydraulic gradients in the aquifer, their salts would gradually be released and lead to the formation of saline groundwater. The formation of saline waters in deeper aquifer sections could lead to CO_2 excess (Equation (1)), as observed in the deep observation well at the Delhi field site.

The elevation of the groundwater table and the leaching of salt stores from the unsaturated zone (Figure 1d) are often discussed as sources of inland salinity in Australia (Herczeg *et al*., 1993; Smith and Compton, 2004; Duncan *et al*., 2005). They may also have played a role in northern India. According to Singh (2005), the installation of the irrigation canals that spread all over Haryana and Delhi

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was often followed by groundwater table elevation and salinity ingress.

For instance, after the extension of the irrigation network in the 19th century, British engineers observed the elevation of water tables that once lay 36 to 39 m below ground to a depth of 4 to 6 m in some villages in and around Delhi (Singh, 2005). Salinity was usually interpreted because of waterlogging, but it was often observed after just a few years (Singh, 2005). Considering that the aquifers were refilled with fresh canal seepage, it seems unlikely that the rapid mineralisation was only caused by evaporation. At least in the initial phase, the dissolution of salts from the unsaturated zone may have contributed significantly to salinity ingress in the shallow groundwater.

Leaching aeolian deposits (e.g. relict dunes; Figure 2) provides an additional source of soluble salts in the study area (Subramanian and Saxena, 1983). Strong winds from the Thar Desert transport large quantities of dust and aerosols, including additions of carbonates, gypsum and halite (Thussu, 2006).

Mixing with saline waters

In Delhi and Haryana, fresh or brackish shallow groundwater is often underlain by saline water (CGWB-GOI, 2006; Rao *et al*., 2007; Kumar *et al*., 2008). This trend could be confirmed by multilevel sampling at the Delhi field site, where salinity gradually increases with depth. The high mineralisation, the relatively low depletion in light isotopes ($\delta^{18}O < -4.5‰$) and the absence of NO_3^- and CO_2 excess are indicators of groundwater from deep sources. They may have formed through by leaching evaporite sediments (e.g. salt clay) as described earlier.

In the study area, highly brackish and saline groundwater was found only around the Najafgarh drain and north of it (Figure 3), so deep saline water seems to be more prevalent in the central parts of the alluvial plains. The southern part, between the drain and the Aravalli range, is dominated by fresh and slightly brackish waters. The thickness of the alluvium decreases towards the south (Thussu, 2006; CGWB-GOI, 2006), and the aquifer seems to be recharged with freshwater from the hard rock range. In Haryana, the Najafgarh drain acts as a hydraulic barrier when surface water seeps to groundwater in the Haryana territory. Downstream the weir at the Delhi field site, temperature gradients (Figure 4) indicate that the channel is actively draining the territory. The hydraulic head anomaly under the discharge area possibly causes the upconing of deeper saline waters (Figure 1e) and mixing with shallow brackish groundwater.

Upconing (Figure 1d) is also a risk below extraction wells and may be responsible for the breakthrough of saline water plumes into the shallow aquifer (Rao *et al*., 2007). Increased mixing may be another consequence of anthropogenic activities in the study area (Datta *et al*., 1996). In the Indo-Gangetic plains of northern India, hydraulic engineering, interbasin water transfer and land use patterns have altered the natural equilibrium of the water cycle since ancient history (Singh, 2005; Ambast *et al*., 2006). For instance, groundwater table elevations of several metres were observed around Delhi, after the extension of an irrigation network in the 19th century (Singh, 2005). In the past few decades, irrigation wells that penetrate unsystematically to different depths also enable mixing processes that would not occur under natural conditions (Seiler *et al*., 2001). Excessive groundwater abstraction has led to a rapid decline of the groundwater table (CGWB-GOI, 2006; Rodell *et al*., 2009). Land use and water resource management must have had a considerable effect on the natural hydraulic system, with consequences such as deeper circulation, increased flow velocities or inversed gradients or upconing. As a consequence of the man-made perturbation of the natural conditions, saline groundwater that was previously preserved in deeper aquifer sections, in stable density stratification, low permeable sediments or hydraulic traps may have been mobilised and transported into the active zone of the meteoric water cycle.

CONCLUSIONS

The study has shown that salinisation of shallow aquifers in the study area is not a homogeneous process but is related to different sources and dynamics, with variations in space and time. There is no indication of a contribution of seawater or the dissolution of evaporates of marine origin.

The accumulation of salts is favoured by the geographical characteristics of the study region, especially the permanent influx of water into a poorly drained sedimentary basin along with a warm and dry climate. In the geologic past, more arid periods may have led to the formation of salt stores in soil, sediments and groundwater. In prehistoric times, the accumulation and flushing of salts occurred in equilibrium with the natural environmental conditions. The man-made alteration of the hydrologic state has triggered the mobilisation of salts and salinity ingress in the semiarid plains. Different processes that may have contributed to salinity and salinity ingress are summarised in Table III. Processes in the first row are those that regulated the water and salt balance in the natural equilibrium. The second line lists potential consequences of land clearance and hundreds of years of canal engineering and flood irrigation. The last line refers to the influence of tube well irrigation, which was the dominating factor during the last decades.

Agriculture in the densely populated plains is essential for the development of the region and the well-being of the rural population. It is highly dependent on irrigation water and threatened by widespread salinity ingress. The results of this study make clear that groundwater monitoring is important for understanding salinity ingress and implementing counteractive measures.

Evaporation during irrigation should be minimised and waterlogging must be prevented to reduce evapotranspiration losses. However, building drainage systems, which are often recommended to prevent waterlogging, creates artificial discharge areas and thereby augments the risk of mobilising deeper saline waters through upconing. Irrigation wells

Table III. Potential sources of groundwater salinity and influences on the salt/water balance through the man-made perturbation of the natural equilibrium

	Evaporation	Formation and leaching of salt stores	Mixing
Natural equilibrium in prehistoric times (adaption of salt/water balance to the natural variation of environmental conditions):	- Temporarily limited: during infiltration of rainwater after rain events - Local hot spots: areas with shallow groundwater tables (e.g. along rivers or in undrained basins)	- Formation of salt deposits in the unsaturated zone - Limited leaching in poorly drained, low-permeability sediments (e.g. salt clay) - Aeolian deposits from the Thar Desert provide an additional source of salts	- Hydraulic equilibrium leads to the formation of density-stratified groundwater - Stable hydraulic conditions limit the leaching of poorly drained sediments
Influences of land use, canal engineering, interbasin water transfer and flood irrigation:	- Evaporation from canal-irrigated fields throughout the year - Groundwater-level elevation along canals enables evaporation through capillary rise in large areas	- Dissolution of salt stores in the unsaturated zone after groundwater table elevation - Changing groundwater dynamics may increase salt leaching from low-permeability sediments (e.g. salt clay)	- Alteration of the hydraulic equilibrium and mobilisation of previously quasi-stagnant saline water - Risk of upconing under drainage canals
Influences since the green revolution and the implementation of excessive groundwater irrigation:	- Area-wide evaporation from groundwater-irrigated fields throughout the year - Recycling irrigation return flow gradually increases salinity - Soil degradation limits seepage and increases evaporation during infiltration	- Changing groundwater dynamics may increase salt leaching from low-permeability sediments (e.g. salt clay)	- High risk of the upconing of saltwater under abstraction wells - Mixing around wells that abstract water from different aquifer sections

should be systematically planned, and management strategies for groundwater irrigation are essential. Any major change in land use and water management can result in a perturbation of the hydraulic equilibrium and should be evaluated as a potential catalyser for salinity ingress.

The results of this study contribute to a better understanding of groundwater dynamics and salinity ingress. This is a key condition for the development of best practice management strategies for integrated water resources in Northern India. Explanatory models developed in other parts of the world could be adapted to the regional context. For similar studies, holistic approaches are recommended, taking into account natural as well as anthropogenic influences on the salt/water balance within a catchment.

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