5 Groundwater flow modelling

The transport of arsenic is determined by the overall movement of groundwater. It is therefore important to understand the pattern of groundwater flow and the factors that may affect it in order to understand the present and possible future movement of arsenic. If groundwater flow rates are extremely small, then diffusion will be the dominant mechanism by which arsenic moves. The diffusive flux is often orders of magnitude lower than advective flux in a flowing system. The transport of arsenic in systems without a significant advective flux will therefore be extremely slow.

The complexity of the Bangladesh aquifers in terms of the variation in sedimentological characteristics both laterally and with depth is clearly very great (Chapter 3). Indeed, it is probably so great that the difficulty and expense of building hydrogeological models of water and solute movement incorporating the level of detail seen in the borehole logs would rarely be justified. Below, the effect that different simplifications of the aquifer system can have on groundwater flow patterns is examined.

In order to understand the movement of arsenic in aquifers, it is important to have some knowledge of the way that water typically moves in the aquifers and how this varies under different flow conditions and management policies – under natural flows and with abstraction for domestic use (hand pumps) and for irrigation (motorised pumps). Some estimate is required of the depth to which groundwater flow occurs and what impact pumping has on the depth of flow. This is particularly relevant for the groundwater arsenic problem since it is known that the highest arsenic concentrations are found in the ‘shallow’ aquifer and there is naturally concern about the possible movement of arsenic to the, as yet, largely uncontaminated ‘deep’ aquifer. We were also interested in the effect that river geometry, specifically meanders, could have on the initial stages of aquifer flushing. Questions such as these can be answered at least in part with simple hydrogeological models. The development of such models is described below.

5.1 OBJECTIVES OF MODELLING

The overall aim was to examine groundwater flow, with the specific objectives as follows:

1. calculate the relative magnitudes of flow in shallow and deep aquifers
2. investigate whether downward flow contaminates the deeper part of the groundwater system with arsenic
3. determine the effect of pumping on the movement of arsenic to irrigation wells and deep tubewells
4. estimate over what timescale arsenic will affect pumping wells.

In addition to these objectives, the regional flow of groundwater is discussed and its possible influence on distribution and movement of arsenic evaluated.

5.1.1 Methodology and approach

The modelling was undertaken in four steps:

1. develop an understanding of how the aquifer system works by creating a vertical slice model and using typical parameters and boundary conditions for Bangladesh aquifers;
2. determine how flow is distributed vertically through the groundwater system using particle tracking techniques;
3. increase the complexity of layering and investigate the effect of this on the vertical distribution of flows;
4. introduce abstraction and determine the travel times from the water table to the well screen.

5.1.2 Simplifications

The purpose of the modelling presented in this section is to investigate the patterns of flow in both the shallow and deep aquifers and what influences them.

In order to investigate these effects, a staged approach was taken, starting with a simple model and then increasing the complexity. It was necessary to simplify the work sufficiently to enable the influence of the hydraulic properties of the aquifer and the effect of pumping to be evaluated. As discussed in Section 4.3, the groundwater system exhibits marked seasonality due to the monsoon rains during May to September. This results in a seasonal pattern of rise and fall of groundwater heads influenced by tidal responses, recharge and increased river stage. The monsoon rains are important in producing seasonality in surface water bodies and groundwater systems but the primary aim of this work was to determine the long-term influence on the vertical distribution of groundwater flows and their effect on arsenic distribution. Therefore, for simplicity, seasonality was not included but it should be included if the models are to be developed further.

5.2 GENERIC MODEL

5.2.1 Introduction

With the exception of the extreme north-west region of Bangladesh, the hydraulic gradients are generally low reflecting the flat topography of the delta region (Section 4.6). However, even small gradients can be important in highly transmissive environments especially over long timescales, and the timescales of interest here vary from a few years up to geological timescales of tens of thousands
of years. The following factors could affect these gradients:

- the balance between recharge and evaporation which affects the water table height;
- the transmissivity and vertical hydraulic conductivity of the aquifer;
- the spacing between rivers;
- the location and depth of the aquifer.

The modelling presented in this Chapter has used MODFLOW and MODPATH to determine the factors that control the vertical distribution of groundwater flow. Vertical flows are determined by the vertical distribution of heads, i.e. hydraulic gradients. These are characterised by:

- the depth of aquifer;
- the amount of recharge;
- the presence of low hydraulic conductivity layers;
- the vertical distribution of hydraulic conductivity.

### 5.2.2 Methodology

The approach adopted was to set up a steady-state two-dimensional vertical-slice model using appropriate parameters, which are described below. Once this had been done, particle tracking was used to trace the movement of the water from the surface through the aquifer. By counting how many particles travelled through a particular horizon, the model could be used to determine the flows at different depths. Flows were calculated for three horizons within the model which were representative of the thicknesses of the 3-layer conceptual model used for the Faridpur region. This conceptual model was introduced in Section 4.2 and further developed in Section 4.9. Flows were calculated for the following three horizons:

1. **Upper shallow aquifer**: 0–40 m below ground level (bgl);
2. **Lower shallow aquifer**: 40–130 m bgl;
3. **Deep aquifer**: 130–400 m bgl.

Once the flow through the different horizons had been established, a number of sensitivity simulations were undertaken to investigate how the patterns of flow changed in response to changing various model parameters. The final application of the model was to introduce a 3-layer hydrogeological model in which each of the layers had a different hydraulic conductivity. This enabled the distribution of vertical flows to be determined based on a more realistic representation of the actual vertical heterogeneity of the aquifers in Bangladesh.

### 5.2.3 Description of the base case

The ‘base case’ here is a steady-state vertical-slice model with a uniform hydraulic conductivity. As discussed above, the steady state assumption ignores the obvious seasonality of the groundwater system. This was necessary in order to simplify the problem.

The main features of the base case model are presented in Figure 5.1 and detailed below:

- thickness 400 m;
- length 2550 m;
- hydraulic conductivity 15 m d\(^{-1}\);
- ratio of horizontal to vertical conductivity 1:10;
- net recharge of 0.3 mm d\(^{-1}\);
- no flow boundaries all around model, with the exception of a single river as the only outflow.

The thickness of the aquifer was chosen to be 400 m to allow the bottom boundary to be deep enough so as not to influence the pattern of groundwater flow.

The hydraulic conductivity is derived from the transmissivity given in the hydrogeological map of the Faridpur district. This indicates a transmissivity of between 2000–3500 m\(^2\) d\(^{-1}\) for an aquifer thickness of 200 m. Thus a hydraulic conductivity of between 10 and 17 m d\(^{-1}\) is suggested, and 15 m d\(^{-1}\) was chosen as a representative value for the present modelling. The ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity is based on the flow modelling undertaken in Phase I (DPHE/BGS/MML, 1999).

No-flow boundaries were chosen for both the base of the aquifer and the lateral boundaries. The lateral boundaries therefore represent the limits of the groundwater catchments. Early calculations with the model demonstrated that the river spacing was important in order to keep the regional gradients to the low values observed in the field. This implies that rivers provide one of the important controls for keeping groundwater gradients closely tied to topography.

The net recharge estimate of 0.3 mm d\(^{-1}\) is also based on the work carried out in Phase I (see Table 4.2 in Volume S3 of DPHE/BGS/MML, 1999). This is summarised in Table 5.1. Although low, this estimate is representative of the difference between the long-term infiltration and the long-term average evaporation.

This value of recharge so derived can be checked by using the observed seasonal variation in groundwater head (see Section 4.6), coupled with an estimate of specific yield to obtain an order of magnitude estimate of recharge. The results from this calculation are presented in Table 5.2 and...
show that recharge could be in the range 0.3 to 5.5 mm d\(^{-1}\). The upper value (5.5 mm d\(^{-1}\)) is thought to be unrealistically high and would be influenced by abstraction. The estimate of recharge used in the modelling study is therefore at the lower end of the likely range. A simple sensitivity analysis was undertaken to address any uncertainty in the value chosen.

The head distribution in the top layer and flow distribution for the base case are presented in Figure 5.2. Examining the gradients (Figure 5.2a) shows that the groundwater gradient to the river is 0.15 m km\(^{-1}\). This is typical of the regional gradients estimated from the latest hydrogeological map for Bangladesh and summarised in Table 4.11.

The 3-layer model used in this part of the investigation is based in part on that used in previous regional modelling studies in Bangladesh and is described in Table 5.3 and Figure 5.3. The top two layers used in this model are equivalent to the top four layers used in previous models. However, this structure is not appropriate everywhere in Bangladesh. For example, there is no well-defined aquitard between the upper and lower aquifers at Faridpur and so this case is treated separately below.

5.2.4 Sensitivity

A series of simulations were undertaken with the aim of testing the sensitivity of the base case results to changes in the various parameters. This was achieved by changing one parameter or feature at a time. The principal results of these simulations are given in Table 5.4.

The main conclusions from the hydrogeological layer model based on a single, uniform layer can be summarised as follows:

- realistic variations in the amount of recharge have little or no effect on the overall pattern of flow;
- changing the thickness of the aquifer makes a small difference to the overall pattern of flow – increasing the

<table>
<thead>
<tr>
<th>Month</th>
<th>Potential recharge</th>
<th>ET</th>
<th>Actual recharge</th>
<th>River in</th>
<th>River out</th>
<th>River net</th>
<th>Change in storage</th>
<th>Balance</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.42</td>
<td>-1.42</td>
<td>-1.00</td>
<td>0.00</td>
<td>-0.96</td>
<td>-0.96</td>
<td>1.96</td>
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<td>-1.38</td>
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<td>-0.75</td>
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<td>-0.01</td>
</tr>
<tr>
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<td>-0.90</td>
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<td>-0.49</td>
<td>1.39</td>
<td>-0.01</td>
</tr>
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<td>-0.51</td>
<td>0.39</td>
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<td>0.19</td>
<td>0.32</td>
<td>0.00</td>
</tr>
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<td>0.70</td>
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<td>-0.13</td>
<td>0.20</td>
<td>-0.90</td>
<td>0.00</td>
</tr>
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<td>2.16</td>
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<td>0.00</td>
<td>0.69</td>
<td>-2.84</td>
<td>0.01</td>
</tr>
<tr>
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<td>3.23</td>
<td>0.53</td>
<td>0.00</td>
<td>0.53</td>
<td>-3.76</td>
<td>0.00</td>
</tr>
<tr>
<td>8</td>
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<td>-3.49</td>
<td>1.16</td>
<td>0.66</td>
<td>0.00</td>
<td>0.66</td>
<td>-1.82</td>
<td>0.00</td>
</tr>
<tr>
<td>9</td>
<td>4.27</td>
<td>-3.69</td>
<td>0.58</td>
<td>0.12</td>
<td>-0.15</td>
<td>-0.02</td>
<td>-0.55</td>
<td>0.00</td>
</tr>
<tr>
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<td>3.35</td>
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<td>-0.20</td>
<td>0.00</td>
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<td>-0.74</td>
<td>0.94</td>
<td>0.00</td>
</tr>
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<td>0.00</td>
</tr>
<tr>
<td>12</td>
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<td>-0.54</td>
<td>0.00</td>
<td>-1.37</td>
<td>-1.37</td>
<td>1.90</td>
<td>0.00</td>
</tr>
<tr>
<td>Average</td>
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<td>-2.10</td>
<td>0.29</td>
<td>0.23</td>
<td>-0.53</td>
<td>-0.30</td>
<td>0.01</td>
<td>0.00</td>
</tr>
</tbody>
</table>

Table 5.2. Recharge estimate based on seasonal groundwater head fluctuations

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Lower value</th>
<th>Upper value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seasonal head variation (m)</td>
<td>2</td>
<td>10</td>
</tr>
<tr>
<td>Specific yield</td>
<td>0.05</td>
<td>0.2</td>
</tr>
<tr>
<td>Recharge (mm d(^{-1}))</td>
<td>5.5</td>
<td>0.3</td>
</tr>
</tbody>
</table>

Figure 5.2. Results from basecase vertical slice model. (a) groundwater head profile; (b) particle tracks.
thickness of the aquifer decreases the flow in the upper
layer;

- introducing a low hydraulic conductivity layer dramati-
cally reduces flow below the depth of this layer.

Introducing the 3-layer hydraulic conductivity model, how-
ever, increases the flow through the middle layer at the
expense of flow through both the upper and lower layers.
The most transmissive part of both the 3-layer and 4-layer
models occurs in the layer representing sediments depos-
ited between the highstand and lowstand limit. This will
therefore be the layer through which the majority of the
flow occurs. For Faridpur, this is between 40 and
130 m bgl, but the exact depth interval will vary in differ-
ent parts of the country.

5.3 SITE SPECIFIC MODEL: FARIDPUR

The aim of this modelling work was to see how the flows
changed with depth as the complexity of the representa-
tion of the layering increased. Four simulations were used
to progress from a homogeneous aquifer to a layered aqui-
fer based on the Faridpur borehole log. The model was
split into three horizons and the vertical distribution of
groundwater flow was determined as described above. The
flow passing though each of these layers was calculated for
each of the four simulations.

5.3.1 Outline geology and hydrogeology

The Special Study Area of Faridpur was chosen as an
example area for determining groundwater flows in a rela-
vitably well-defined hydrogeological setting. The geology of
the Faridpur area is described in Section 3.3.4. The general
geological features of the area are:

- the *upazila* is crossed by a recently-incised river channel
  which was active at the last glacial maximum (about
  15 ka to 20 ka ago);
- this channel which is about 4 km wide sits on top of a
  12 km main channel which contains meanders which
  predate the last interglacial more than 120 ka ago;
- subsidence since the main channel was formed has led
  to the stacking of channel deposits. There has been
  subsidence of about 60 m during the last 120 ka at a
  rate of about 0.5 mm a⁻¹ (Figure 3.12);
- the *upazila* lies in a possible subduction zone – the
  Faridpur trough which runs north-west towards the
  Sylhet Basin.

The sediments consist of two partly upwardly-fining
sequences. The lower one is a transgressive tract which
fines from coarse sands and gravels to medium to fine
sands. The upper one shows a sudden change in lithology
from silt to fine to medium sands and finer. This could be
caused by channel switching with the recently-incised
channel being a distributary of the main channel, or the
smaller channel may have been captured by the main chan-

### 5.3.2 Methodology

The simple vertical-slice model described in Section 5.2
was modified to incorporate a more realistic representation
of the geology as observed in the borehole log of the
Faridpur cored borehole (Figure 3.13). This log showed
that there was a high degree of layering within the aquifer.
The complexity of the model was therefore increased from
the homogeneous aquifer through a simple 3-layer block
model to one based more closely on the actual geological
log of the borehole. This progression from homogeneous
model to a complex layered model is illustrated in Figure
5.3. The model had the following features and assump-

- steady state model;
- the River Kumar was assumed to be the only outflow;
- recharge was assumed to be 0.3 mm d⁻¹ over the whole
  area;
- the initial hydraulic conductivity was 15 m d⁻¹;
- the aquifer was 400 m thick.

Starting with a uniform hydraulic conductivity model, the
layering within the model was developed through three
further simulations:

- VS1: uniform hydraulic conductivity;
- VS2: 3-layer model; transmissivity the same as VS1 (see
  Table 5.5) with the hydraulic conductivities representa-
  tive of the appropriate values for each hydrogeological
  layer; fine sand/silt for layer 1, medium sand for layer 2
  and silty clay for layer 3;
- VS3: 3-layer model using hydraulic conductivities based
  on the observed and simplified lithologies of each layer
  with standard values for each lithology based on per-
  meameter tests and pumping test analysis (Barker et al.,
  1989);
- VS4: 16-layer model derived from the detailed borehole
  log and typical hydraulic conductivity values (see Table
  4.10 in Section 4.4) used for VS3 (see Table 5.6).

### 5.3.3 Effect of increasing complexity

The flow through the three different horizons was
assessed by counting the number of particles that passed
through a particular horizon. These horizons are based on
the 3-layer model presented in Table 5.5. Accounting for
the flows in this way meant that all the particles travelled
through the top horizon (100%), but that decreasing num-
bers of particles were tracked through the middle and
lower horizons. The percentage of the total flow associated
with each horizon was then calculated from the proportion
of particles passing through each horizon. The results for the four simulations (VS1 to VS4) are given in Table 5.7.

The results show that for the uniform hydraulic conductivity example (VS1) the upper and lower shallow aquifer layers receive one half and one quarter of the flow, respectively (Table 5.7). The basic 3-layer model (VS2) defines the upper layer as significantly less permeable than the middle layer (Table 5.7). This is reflected in an increase in flow to the middle layer from 52% for the uniform hydraulic conductivity model to 64% in the 3-layer model. In contrast, relatively little flow (<10%) reaches the deeper part of the system, the Pleistocene deposits, in this simulation. This flow pattern is confirmed by the 3-layer model using the hydraulic conductivities estimated from the simplified lithology used in earlier work (VS3). Table 5.7 shows that the main difference is that less flow reaches the deeper aquifer, half as much as in the uniform aquifer model.

When the layering is based on the more complex geology derived from the Faridpur borehole log, there is no significant change in the pattern of flow to the main aquifer (Table 5.7) but flow to the deeper part of the system has increased from 4% to 12%. This is due to the sequence of coarse sand at 110–134 m (Figure 3.13). The particle tracking shows that significant flow occurs through this layer in contrast to that based on the simpler 3-layer models. This high hydraulic conductivity layer draws water deeper into the aquifer as illustrated in Figure 5.4.

Additional simulations were undertaken to address the uncertainty in the recharge rate and the role of the coarse sand layer. Doubling or halving the recharge rate had little or no effect on the overall flow pattern, but increasing the recharge rate did increase the depth to which flow occurred. Similarly, doubling or halving the thickness of the coarse sand layer produced little or no difference to the overall flow pattern.

A conclusion from these simulations is that, perhaps not surprisingly, the way that the inherent stratification in the aquifer is represented is important, particularly with regard to the possible flows to the deep aquifer. Since the VS4 simulation comes closest to geological reality, then the assumption of a uniform hydraulic conductivity (VS1) is clearly a poor representation of the flow through the system as a whole. The remaining representations all give a similar flow through the middle layer, the main aquifer, but the amount of flow to the deeper aquifer varies significantly.

While all simulations predict that only a small percentage of the overall flow passes through the deep aquifer under natural flow conditions, this varies from 4% to 12%, a factor of three. Therefore although flow through the deep aquifer is relatively small, it could be significant in the long term as these sediments will have been subject to groundwater flow for a long period of time, e.g. many thousands of years.

It would also be expected that the converse would apply. Low hydraulic conductivity layers should have the opposite effect to high hydraulic conductivity layers. This is confirmed by work undertaken with the generic model presented in Section 5.2. This demonstrated that relatively impermeable horizons will significantly reduce flow to deeper horizons. In practice, these low hydraulic conductivity horizons are not laterally extensive and this will also impact on the overall flow patterns. In particular, groundwater flow will tend to bypass localised low permeability horizons. This is significant because of the generally high positive correlation between the arsenic content and texture, i.e. fine-grained horizons with a low permeability tend to contain more arsenic than coarse-grained horizons (Chapter 11). Therefore, where discrete, these high-arsenic zones will tend to be protected to some extent from flushing.

### 5.3.4 Effect of abstraction

The work presented above has examined the patterns of groundwater flow under natural flow conditions. However, there is obviously concern that a presently uncontaminated well may become contaminated with arsenic as a result of the movement of groundwater induced by pumping. Clearly the likelihood of this depends on the distribution

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (m)</th>
<th>Flow from surface through layer</th>
<th>Flow with deepest extent in layer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper shallow</td>
<td>0–40</td>
<td>100 100 100 100</td>
<td>VS1 VS2 VS3 VS4 VS1 VS2 VS3 VS4</td>
</tr>
<tr>
<td>Lower shallow</td>
<td>40–130</td>
<td>52 64 60 68</td>
<td>48 36 40 32</td>
</tr>
<tr>
<td>Deep</td>
<td>&gt;130</td>
<td>24 8 4 12</td>
<td>24 8 4 12</td>
</tr>
</tbody>
</table>

**Figure 5.4.** Flowlines for the base case model. Flow travels from the left to a discharge point on the right.
of arsenic within the aquifer and the nature of the groundwater flow induced by pumping. This itself depends on the nature of groundwater abstraction.

As discussed in Chapter 4, groundwater abstraction in Bangladesh is based on a number of different technologies:

• hand pump tubewells (HTWs) for village supply purposes;
• shallow tubewells (STWs) with motorised pumps for irrigation;
• deep tubewells (DTWs) with motorised pumps, also for irrigation;
• public water supply (PWS) wells used for town and city water supply.

Each type of well has a different abstraction pattern over time. Hand-pump tubewells and PWS wells operate throughout the year, whereas irrigation wells are used for 2–3 months during the dry season and only when required by the climatic conditions. In order to determine the likely effect of abstraction on groundwater flow, a long-term average abstraction rate was used in conjunction with a steady state groundwater model. This is a simple first step in assessing the possible impact of abstraction on the vertical distribution of groundwater flows.

Annual average abstraction was estimated and applied to the VS4 case used for the Faridpur vertical-slice model. The occasional use of tubewells for irrigation in Bangladesh means that it is difficult to obtain accurate estimates of abstraction rates by irrigation wells. For the purposes of the modelling described in this section, these were obtained from an estimate of the regional coverage of wells and their individual abstraction rates (DPHE/BGS/MML, 1999). These were then converted to give a combined outflow for the modelled area. Sixteen shallow tubewells (STWs) and 2 deep tubewells (DTWs) were used to represent the pumping from the modelled area.

The total abstraction was based on an equivalent recharge calculation using the total for the whole of Faridpur upazila (55.5 Mm³ a⁻¹ for STWs and 6.5 Mm³ a⁻¹ for DTWs).

There are now believed to be some 6–11 million hand-pump tubewells in Bangladesh. Although, in contrast to irrigation wells, these are operated throughout the year for domestic water supply, many are only for a single household. Individually, therefore they use very low volumes of water, mainly from the upper part of the shallow aquifer, and are widely distributed throughout the country. While the total groundwater used for domestic purposes may be significant, the impact upon the groundwater system is likely to be small and is not taken into account in the model. Table 5.8 gives details of the depths and assumed flow rates for each set of wells. In summary, the following conditions have been assumed:

• the overall abstraction rate was made equivalent to the estimated average for Faridpur upazila (62 Mm³ a⁻¹);
• typical irrigation abstractions for shallow tubewells (STW) and ‘deep’ tubewells (DTW) together with public water supply (PWS) boreholes were assumed (Table 5.8);
• STWs were set at 65–75 m depth;
• DTWs and PWS wells were set at 110–130 m depth.

Note that the distinction between ‘shallow’ and ‘deep’ tubewells used here is based on that used in previous drilling and well construction programmes undertaken in Bangladesh. In Faridpur, the ‘deep’ aquifer is at a shallower depth than in other parts of Bangladesh and can also be called the lower shallow aquifer since it is not separated from the upper shallow aquifer by a thick clay layer.

Not surprisingly, introducing abstraction markedly changes the pattern of groundwater flow compared with the natural flow pattern. Groundwater flow in the pumped system is illustrated in Figures 5.5 and 5.6 and should be compared with the natural groundwater flow pattern in Figure 5.4. Figure 5.5 demonstrates the pattern of flow to the STWs and Figure 5.6 illustrates the groundwater flow to the DTWs from the river.

Comparing the simulations with and without abstraction demonstrates that with abstraction:

• with abstraction, water flows from the river to the aquifer;
• there is a reversal of the local hydraulic gradient compared with the natural base case;
• there is an increase in vertical gradients;

<table>
<thead>
<tr>
<th>Well</th>
<th>Depth of well screen (m)</th>
<th>Total modelled abstraction (m³ d⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>STW</td>
<td>65 – 75</td>
<td>100</td>
</tr>
<tr>
<td>DTW &amp; PWS</td>
<td>110 – 135</td>
<td>15</td>
</tr>
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</table>

Total abstraction is based on an equivalent recharge calculation using the total for Faridpur upazila (55.5 Mm³ a⁻¹ for STWs and 6.5 Mm³ a⁻¹ for DTWs).

Figure 5.5. Flowlines from the surface to STWs for the base case when pumping is included. Flow travels downwards.
the flow paths are very steep vertically and the lateral extent from each STW is small (about 100 m) implying a narrow distribution of travel times.

In the scenario outlined above, flow to the wells requires more water than is available from recharge when the same rate of recharge is used as in the unpumped model. This leads to the local gradient of the groundwater reversing. Groundwater is drawn from the rivers, i.e. the river becomes influent and acts as a recharge zone rather than a discharge zone. In practice, there will be some additional recharge from the irrigation returns.

In order to illustrate this effect in more detail, the results of two model runs are given below assuming either ‘average’ recharge rates (0.3 mm d⁻¹) or ‘high’ recharge rates (0.6 mm d⁻¹). Table 5.9 shows that, as before, the flow to the wells is predominantly through the middle layer but that with pumping a greater percentage of the recharge passes through the middle layer especially under the average recharge conditions. In the average recharge scenario, flow through the middle layer accounts for all of the natural recharge at the surface as well as any induced recharge from the river to the aquifer. This has the effect of reducing flow to the deeper part of the groundwater system either completely, as in the average recharge case, or to an extremely small value as in the high recharge case. In the high recharge case, abstraction produces the following effects:

• water now flows from the aquifer to the river, i.e. there is no longer any induced recharge from the river. Recharge is greater than abstraction;

• pumping still causes a reduction in the regional hydraulic gradient;

• there is an increased flow to the middle layer at the expense of both the upper and lower layers.

Placing the wells in the middle layer has reduced flow through the other layers and in this respect the wells act as interceptor wells, protecting the deeper aquifer from any contamination that might be derived from the shallower horizons.

It is possible to use particle tracking techniques to estimate the time of travel of groundwater from the water table to the well screen. Assuming a porosity of 20%, the resulting travel times vary from 40 to 300 years (Table 5.10). The contrast between the unpumped and pumped situations is illustrated in Table 5.11. This demonstrates that 50% of the flow to the river takes more than 225 years, whereas 50% of the flow to the STWs takes either 86 years for the average recharge case or 47 years for the high recharge case.

This situation is reversed for the time taken for less

<table>
<thead>
<tr>
<th>Table 5.9. Distribution of flow in the aquifer under natural and pumped conditions</th>
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<tbody>
<tr>
<td><strong>Percentage of overall flow at given depths</strong></td>
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<tr>
<td><strong>Flow from surface through layer</strong></td>
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<tr>
<td><strong>Layer</strong></td>
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<tr>
<td>Upper shallow</td>
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<tr>
<td>Lower shallow</td>
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<tr>
<td>Deep</td>
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| Figure 5.6. Flowlines to a DTW for the basecase model with pumping included. |

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<thead>
<tr>
<th>Table 5.10. Approximate times of travel from the water table in the well catchment area to the various targets with pumping</th>
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<tr>
<td><strong>Target</strong></td>
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<tr>
<td><strong>Average recharge</strong></td>
</tr>
<tr>
<td>River</td>
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<tr>
<td>STW</td>
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<td>DTW/PWS</td>
</tr>
</tbody>
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<tr>
<th>Table 5.11. Distribution of flows by time of travel from the water table to the well screen</th>
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<tr>
<td><strong>Percentage of flow</strong></td>
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<tr>
<td><strong>Time taken for groundwater to flow from the water table to the discharge point (years)</strong></td>
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<tr>
<td><strong>Natural River</strong></td>
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<tr>
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than 95% of the flow to reach a particular target. For the unpumped case, less than 95% of the flow takes 28 years to move from the water table to the well screen. This compares with 55 and 41 years under average and high recharge conditions, respectively, for the STWs. This is a direct result of the geometry of the two situations. The distance between the water table and the river is smaller near the river than the distance between the water table and the STWs.

For the upper shallow aquifer, assuming well screens at 65–75 m below the water table, it was estimated that 50% of the flow from the water table took less than 50 years to reach the well (Table 5.11). However, this is highly dependent on the recharge rate: the higher the rate, the shorter the travel time. The approximate maximum lateral distance of flow from the water table to the STWs was estimated to be around 50–125 m.

For the lower shallow aquifer, assuming a well screen at 110–135 m below the water table, the travel time under pumped conditions was estimated to be in excess of 200 years from a lateral distance of approximately 500 m. Under natural (unpumped) conditions, flow to the same depths was estimated to take in excess of 300 years, with a lateral movement of 1000 m. These travel-time estimates are consistent with the observed presence of tritium in the upper part of the shallow aquifer and its absence from the deep aquifer (see Chapter 7).

These travel times indicate that presently uncontaminated wells could start to show evidence of contamination over a timescale of decades. In practice, the actual timescale will also depend strongly on the distribution of arsenic and the retardation due to sorption processes (see Chapter 12).

The overall conclusion is that irrigation wells with realistic abstraction rates significantly affect the flow patterns within the groundwater system. They tend to draw water more quickly through the main aquifer. In this way, the wells are acting as interceptor wells and could protect the deeper parts of the aquifer system from arsenic contamination, particularly under average recharge conditions. The depth of placement of the well screen is therefore an important factor controlling the movement of potentially contaminated water.

5.4 GROUNDWATER FLOW NEAR TO A MEANDERING RIVER

5.4.1 Background to the problem and objectives

Rivers of various sizes are extremely common in Bangladesh. Therefore the interactions between rivers and groundwater are likely to have an important influence on local groundwater flow and ultimately on the arsenic distribution.

In our Phase I report (DPHE/BGS/MML, 1999), a modelling study of the Chapai Nawabganj Special Study Area showed an area of low groundwater flow velocities in the locality of Nawabganj town. This town is situated inside a meander of the River Mahananda and is the location of unusually high groundwater arsenic concentrations. A maximum arsenic concentration of 2400 µg L⁻¹ was found in the town during the present project and the average arsenic concentration found inside the meander was 455 µg L⁻¹. This compares with an average concentration of 144 µg L⁻¹ for the upazila as a whole.

Several explanations can be proposed for such a highly variable distribution of arsenic in this upazila. One hypothesis focuses on variations in the concentration of arsenic present in the original sediments (the source term), another focuses on variations in the processes leading to the mobilisation of arsenic from the sediments, e.g. variations in redox conditions, while a third focuses on the post-depositional flushing of arsenic from the aquifer. The reality of course is that all three sources of variation are likely to be important, although to differing extents in different circumstances. Here, we examine factors affecting the third hypothesis, namely the degree of flushing of the aquifer. In particular, we investigate the impact of variations in river geometry on groundwater movement and consequently on the degree of flushing. This has been inspired in part at least by the Chapai Nawabganj ‘hot spot’ but also by the known occurrence of some other hot spots near major rivers. A cluster of high arsenic wells was also found close to the Wabda river in the village survey of Mandari (see Chapter 8).

The aim of the work, therefore, was to examine groundwater flow around a river meander, and to determine to what extent this influences the rates of groundwater movement close to the meander.

5.4.2 Water movement near to a river meander

Two series of models were developed for this study:

- a generic model – developed to investigate groundwater flow patterns around an idealised river;
- a site-specific model – used to investigate the distribution of groundwater velocities in the vicinity of the River Mahananda as it flows around Chapai Town.

Both of these models were single-layer, steady-state groundwater models. In order to make the transition between the simple, generic model and the site-specific model based on the hydrogeology of Nawabganj upazila, the following steps were undertaken:

- increase the complexity of the distribution of transmissivity;
- create a distribution of recharge;
- impose a gradient in river stage;
- reproduce the shape of the river system.

Simplifications made include the assumption that the Barind Tract could be treated as a vertically homogeneous layer, and the omission of streams flowing across the Barind Tract. The rivers were modelled as having constant stage values whereas in reality, seasonal variations in river stage and groundwater levels are also important in determining groundwater flow. River flows and levels, together with groundwater levels, are highly seasonal, driven by the monsoon rains and tides.

The main aim, however, of this exercise was to investigate the possible influence of river meanders on local groundwater flow. It was necessary, therefore, to simplify...
the problem significantly. Seasonal changes in both river and groundwater flows were therefore not included in the models described below. This has the advantage that the effects of the geometry of river meanders could be thoroughly investigated without the additional complications of seasonal changes in river and groundwater levels and their interactions.

5.4.3 Generic meander model

Methodology

The generic model was developed as follows:

- circular meanders, no river stage gradient (PM1);
- circular meanders, river stage gradient (PM2);
- elliptical meanders, river stage gradient (PM3).

For each model, the groundwater gradient perpendicular to the river was examined. Particle tracking was also used to determine the pattern of flow close to the river.

Description of the initial model

A simple model was first developed to provide a baseline groundwater velocity distribution without having to deal with the complexities of building a detailed model of the Chapai Nawabganj area. This simple model consisted of a single layer with constant aquifer geometry and property values. The boundaries were set as no-flow to avoid the complication of any regional flow. All the recharge therefore flowed to the river, which was the only outflow. The position of the river and boundaries are illustrated in Figure 5.7.

The aquifer geometry and properties were based on those used for the Phase I modelling study of the Chapai Nawabganj area (DPHE/BGS/MML, 1999). The aquifer was assumed to be 50 m thick, had an annual average recharge (taking into account the effects of evapotranspiration and surface runoff) of 0.1 mm d\(^{-1}\), and a hydraulic conductivity of 5 m d\(^{-1}\). The low value of recharge results from the small difference between long-term average infiltration and long-term average evaporation. A river with three semi-circular meanders was positioned with the river axis running north-south halfway across the model grid. In this way, an equal amount of recharge was input on each side of the river.

Results from the generic meander model (PM1)

Attention was focused on the central meander as this was assumed to be independent of boundary effects. The head contours were found to be markedly different on the inside and outside of this meander, with a much wider spacing on the inside of the meander. This observation was quantified by plotting west-east profiles of head gradient, in particular a profile across the apex of the central meander (Figure 5.8). This showed that the groundwater velocity outside the meander, which ranged from 5 mm d\(^{-1}\) to 20 mm d\(^{-1}\) was up to 4 times greater than the velocity inside the meander, 1 mm d\(^{-1}\) to 5 mm d\(^{-1}\). These results demonstrate that groundwater flow velocities are significantly lower inside the river meander. This is further illustrated by a velocity plot (Figure 5.9).

Particle tracking was used as a form of flow visualisation to determine the distribution of flows around the meander. This also illustrated the differing nature of flow inside and outside the meander. Groundwater flow from outside the meander converged towards the apex of the river, whereas flow from the inside of the meander was divergent (Figure 5.10). Particle counting showed that 90% of the baseflow into this stretch of river originated from the outside of the meander.

Effect of introducing gradient in river stage (PM2)

The previous model, PM1, was further developed by putting a gradient on the stage of the river so that the river sloped from north to south. Although this did affect
heads, there was little change in the velocity distribution. The low velocity area inside the meander was reduced in size, however, being confined more to the northern (upstream) part of the meander (Figure 5.11).

**Effect of changing meander shape (PM3)**

This run (PM3) was based on the basic model described above (PM1), but the river meanders were made more oval in shape so that they more closely resembled the actual river geometry. However, the river remained symmetrical about its axis, which was again positioned halfway across the model. As well as more closely resembling the true situation, this model enabled the effect of meander shape to be assessed by comparison with the previous results.

Compared to the results from the basic model (PM1), the head gradients were lower on both sides of the river in simulation PM3. This is because there were more river cells for groundwater to receive the discharge, but the difference in gradients on each side was more marked. The groundwater velocity on the outside of the meander ranged from 2.5 mm d\(^{-1}\) to 10 mm d\(^{-1}\), while on the inside of the meander it was 0.5 mm d\(^{-1}\) to 2 mm d\(^{-1}\), i.e. a factor of five different. These results indicate that the curvature of the meanders controls their impact on local groundwater flow velocities.

Particle tracking showed the same pattern of convergent flow from the outside of the meander to the river with divergent flow inside the meander. The flow to the apex of the river meander from each side of the river was estimated and again showed that 90% of baseflow to the river came from the outside of the meander.

5.4.4 Site-specific meander model: Chapai Nawabganj

The geology of Chapai Nawabganj has been summarised in Section 3.3.3.

**The Chapai Nawabganj meander model (PM4)**

The geology of the Chapai Nawabganj area was simplified for modelling purposes. The upper aquifer was assumed to be a single horizontal layer with a uniform thickness of 50 m and an elevation of 20 m above MSL. The cell dimensions (500 m by 500 m), grid size (60 cells by 65 cells), and the positions of no-flow cells (around the edge
of the model) and river cells, were based on the previous model of the area (DPHE/BGS/MML, 1999), and are shown in Figure 5.12. A no-flow boundary was chosen for the same reasons detailed in Section 5.3. The steps in river stage that were present in the Phase I model were smoothed to provide a more realistic representation of the river.

The transmissivity (T) was defined in three zones (Figure 5.12). An average value was used for the central floodplain area, which is dominated by alluvial deposits associated with the River Mahananda. This was determined from the lithological description given in the log of the project test borehole in the study area combined with the associated hydraulic parameters. The Mahananda floodplain was classified as Zone 1 with \( T = 500 \text{ m}^2 \text{ d}^{-1} \). The Barind Tract, to the east, is made up from less permeable strata and the transmissivity was assumed to be an order of magnitude smaller than that of the Mahananda floodplain. The T of this Zone 2 was set to 50 \( \text{ m}^2 \text{ d}^{-1} \). The western area (Zone 3) made up from thicker and more highly permeable River Ganges deposits was given a transmissivity value of twice that of Zone 1, i.e. 1000 \( \text{ m}^2 \text{ d}^{-1} \).

Initially, the recharge over the entire model was set equal to that used in the simple models, namely an annual-averaged value of 0.1 mm d\(^{-1}\). However, this produced unrealistically high head gradients over the Barind Tract. Further examination of the geology and topography of the Barind Tract showed that very little of the rainfall ever enters the underlying strata due to the extremely low permeability of the surface deposits. Furthermore, the topography slopes to the east in the Barind tract, so most of the runoff will not enter the Mahananda/Ganges floodplain. The T of this Zone 2 was set to 50 \( \text{ m}^2 \text{ d}^{-1} \). The western area (Zone 3) made up from thicker and more highly permeable River Ganges deposits was given a transmissivity value of twice that of Zone 1, i.e. 1000 \( \text{ m}^2 \text{ d}^{-1} \).

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### Results

The groundwater flow velocity distribution obtained from this model (PM4) is shown in Figure 5.13. The effects of a meandering river on groundwater velocity that were observed in the simple models (PM1 to 3) described above are also shown in this model. The groundwater velocity outside the meander was 10 mm d\(^{-1}\) compared with a velocity of 0.6 mm d\(^{-1}\) on the inside the meander. As expected, the low velocity area inside the central meander, where the high arsenic concentrations have been observed, is focused on the northern side due to the gradient in river stage.

However, this area of low velocity groundwater flow within the meander does not correspond exactly with the observed distribution of arsenic in the area and so there are clearly other factors involved – for example, the variation of iron oxide content of the sediments could also be a significant factor. Besides the possible factors mentioned above, the instability of the river systems means that the position of meanders will change with time. On a timescale of thousands of years, this will tend to even out the effects described above and is one factor which might explain why there is not a simple correlation between present-day river geometry and groundwater arsenic concentrations. It may be that the areas with high arsenic concentrations are found where the river systems have been particularly stable.

Nevertheless, the impact that river geometry might have on local flow velocities – a factor of five was estimated here – could be significant given the relatively young age of many Bangladesh sediments in relation to the estimated flushing times of the aquifers (Chapter 4). In more mature aquifer systems, these differences will become less significant.
Model runs were also undertaken with PM4 to assess the sensitivity of groundwater head to changes in various parameters. Figure 5.14 shows that varying recharge and transmissivity (between 50% and 200% of the original values) had very little effect on heads near the rivers, with the effect tending to decrease close to the rivers. This was expected as the stage in the river cells was a constant, hence the river controls the head in its vicinity.

Since the area of high arsenic groundwaters in Chapai Nawabganj is on the inside of the meander and therefore close to a river, groundwater heads and flows in these areas are not sensitive to changes in recharge or transmissivity. These findings do not affect the conclusions reached, and therefore uncertainties in recharge and transmissivity are not of concern. However, as the areas of the model furthest from the river (most notably the Barind Tract) are most sensitive to changes in these parameters, improvements to the accuracy of the model in these areas would need a more detailed assessment of the appropriate values for these parameters.

The sensitivity of simulated heads to changes in the river gradient was also evaluated. Steepening the river gradient caused the low velocity zone to shrink, becoming more restricted in the northern part of the meander. A shallower river gradient had the opposite effect (Figure 5.14).
The results of the sensitivity analyses were also normalised by dividing the change in head ($\Delta h$) by the change in parameter value ($\Delta P$) expressed as a fraction of the original value ($\Delta P/P$) (Figure 5.15). This normalisation enabled the response to changes in different parameters to be compared directly. In general, the head was considerably more sensitive to changes in river gradient than to changes in either recharge or transmissivity. For example, a $\Delta P/P$ of $-0.5$ (halving the parameter value) in river gradient had 3–7 times more impact on the head in a cell in the area around Chapai Town than the same relative change in recharge or transmissivity. However, the same change in parameter values has a different effect in a cell in the north-west of the area where the change in river gradient had the least impact. Recharge there had over 16 times more effect and transmissivity had over 7 times the effect.

5.5 SUMMARY AND CONCLUSIONS

5.5.1 Vertical-slice model

Investigations using the generic vertical-slice model shows that in order to achieve the low groundwater gradients observed, the aquifers have to have a high transmissivity coupled with a low recharge. Rivers also need to be spaced regularly, to provide outflows to the groundwater system. Sensitivity analyses showed that:

- realistic variations in the amount of recharge have little or no effect on the overall pattern of groundwater flow;
- changing the thickness of the aquifer makes a small difference to the overall pattern of flow; increasing the thickness of the aquifer decreases the flow in the upper layer;
- introducing a low hydraulic conductivity layer dramatically reduces flow below this layer.

Three modelling scenarios were investigated: homogeneous (single-layer) model, 3-layer model and a complex layered model based on the BGS Faridpur test borehole log (16 layers). It was found that increasing the complexity of the hydrogeological layering changed the path of flow to the deep aquifer markedly. It decreased the amount of flow from recharge to the river via the deep aquifer from 24% for the homogenous case to 4% for the 3-layer model. Adopting the 16-layer discretisation increases the flow to the deeper part of the system to 12%, an increase by a factor of three.

There is therefore a significant variation in flow pattern depending on the conceptual model of the system chosen. Therefore in order to characterise the flows in the system with greater precision will require a good understanding of the layering within the aquifer system, and any lateral variation. This may ultimately limit the confidence that can be placed on groundwater flow estimates to the deep aquifer.

When the STWs and DTWs typical of those used for irrigation were put into the vertical-slice model, the following were estimated for the pumped flow conditions:

- for the upper shallow aquifer, assuming well screens placed 65–75 m below the water table, it was estimated that 50% of the flow took less than 50 years to reach the well from the surface;
- for the lower shallow aquifer, assuming a well screen at 110–135 m below the water table, the travel time to the well screen under pumped conditions was estimated to be in excess of 200 years from a lateral distance of approximately 500 m.

These results are highly dependent on the recharge rate: the higher the rate, the shorter the travel time. The main implications for arsenic contamination at wells are:

- Irrigation wells with realistic abstraction rates significantly affect the flow patterns within the groundwater system;
- significant flow from surface layers could arrive at wells could over the time-scale of decades;
- retardation due to sorption will be an important mechanism controlling the rate at which any changes in arsenic concentration take place.

The wells draw water more quickly from the shallow aquifer. This has two important consequences. Wells positioned in the shallow aquifer act as interceptor wells and could protect the deeper parts of the system from arsenic contamination. However, siting wells in the deep aquifer will eventually draw water from the more contaminated shallow aquifer and increase arsenic concentrations in the deeper part of the system, though not necessarily significantly. It is not yet clear over what timescale these effects will take place.

5.5.2 Groundwater flow near to river meander

Modelling studies have demonstrated the effect of a river meander on local groundwater flow patterns. Groundwater velocities were estimated to vary from 1 mm d$^{-1}$ on the inside of a meander to 5 mm d$^{-1}$ on the outside of the meander. When a specific river geometry was introduced, based on the River Mahananda at Chapai Nawabganj, the velocity variation ranged from 0.6 mm d$^{-1}$ to 10 mm d$^{-1}$. The effect on groundwater velocity was not limited to meanders but was also observed when an area of the aquifer was surrounded by a river on three sides. Being at least partially surrounded by a river imposes a low hydraulic gradient in the interior – a kind of ‘moat’ effect.

While we have concentrated here on the role of meanders in controlling the local rates of water movement, this is only one factor in controlling the overall distribution of arsenic in groundwater. There are clearly others which also have an important role. For example, there may be accumulations of fine-grained material, including iron oxides, on the insides of meanders and these could lead to a greater-than-average source of arsenic in that region.