12.1 INTRODUCTION

The chapters in this book document sources of spatial variability of hydrologic fluxes and moisture storage in catchments. Some variability reflects that of the atmospheric forcing, and some results from the interaction of that forcing with spatially varying soil, vegetation, and topographic properties. Part of the variability due to interaction, for example that arising from the dependence of soil moisture on soil hydraulic properties during infiltration, may change from storm to storm depending on storm intensity and duration (e.g. Salvucci, 1998). Patterns that arise from the spatial distribution of groundwater–vadose zone interactions, however, have a persistent nature due to the long timescales of groundwater redistribution (see, e.g., Tóth, 1966). The existence, nature and cause of these patterns in the Trochu catchment of Alberta, Canada, are the focus of this chapter.

Throughout the chapter it is assumed that the dominant mode of this interaction (with respect to impact on spatial patterns) is the dependence of surface fluxes on the position of the water table relative to the land surface. This assumption is explored by coupling an equilibrium model (Salvucci and Entekhabi, 1995) that estimates long-term average water table dependent surface fluxes to a groundwater flow model, and then comparing the model results with patterns of recharge and discharge measured by Tóth (1966). The equilibrium model is particularly well suited for comparison with Tóth’s measurements because the latter are largely based on field observations of natural time-integrators of subsurface flow conditions (e.g. presence of salt precipitates). These measurements are a mix of quantitative and observational indicators that provide a spatial picture of long-term recharge and discharge locations.

The water table position relative to the ground surface is assumed to represent the dominant mode of interaction because it impacts on the partitioning of rainfall in two important ways: 1) by bounding the moisture profile, and 2) by
creating a potential source of capillary rise to the root zone. In areas of shallow water table, the bounding of the moisture profile promotes runoff (Dunne and Black, 1970b) and the potential for continuous capillary rise (Gardner, 1958) maintains evapotranspiration at potential rates long after other parts of the landscape dry out. The position of the water table, in turn, depends on the spatial distribution of recharge, capillary rise, and surface water contacts (e.g. springs and lakes). This interdependence of vadose zone and groundwater flows can be viewed either as a consequence of coupled soil moisture and groundwater dynamics, or simply as a constraint imposed by mass conservation; that is, averaged over many wetting and drying cycles, the divergence of the groundwater flow field, which depends on the groundwater pressure distribution and is thus reflected in the water table topography, must balance the net of inputs from and losses to the vadose zone.

12.1.1 Relation to Previous Studies

The central theme of this chapter, the continuity and interdependence of groundwater and vadose zone flows, was recognised and explored in a series of papers by Tóth (1962, 1963, and 1966). Therein a comprehensive analysis of the spatial structure of recharge is detailed, and a theory of regional groundwater flow that accounts for losses through the vadose zone as discharge from the aquifer system is developed. In the third paper, Tóth describes how chemical, biological and piezometric observations can be used to map the spatial distribution of aquifer recharge and discharge areas. In the first two papers he provides methods by which the groundwater flow equation can be solved, for a given fixed water table, in order to predict these patterns.

Despite the recognition of this interdependence three decades ago, many models today either treat groundwater and vadose zone flow systems in isolation, or at most treat conditions at the boundaries between them as fixed quantities. When interactions at the boundaries are ignored, however, potentially important feedbacks are not allowed to occur. For example:

1. Vadose zone analyses that assume a condition of gravity drainage at the bottom of a soil column (e.g. Milly and Eagleson, 1987) may predict recharge to groundwater in excess of what the underlying aquifer can transmit;
2. Climate models that incorporate one-dimensional land surface parameterisations (e.g. Rosenzweig and Abramopoulos, 1997) and ignore lateral groundwater redistribution may fail to simulate large low-lying areas where moisture is evaporated long after higher areas dry out, thus underestimating evaporation and overestimating the sensitivity of evaporation to model parameters;
3. Groundwater studies that take vadose zone inputs as independent of the groundwater flow regime (e.g. Danskin, 1988) can predict artificially high water tables, and those that fix the water table a priori and diagnose
vadose zone fluxes from the groundwater divergence field (e.g. Stoertz and Bradbury, 1989; Ophori and Tóth, 1989) may predict recharge in excess of annual precipitation; and

4. Catchment models that account for near-surface lateral saturated flow (e.g. Beven and Kirkby, 1979; Hatton et al., 1995), but fix the hydraulic gradient to reflect surface topography, constrain the spatial variability of lateral redistribution to be accounted for solely by changes in transmissivity. These models cannot account for the regional groundwater circulations that maintain riparian zones (which in turn affect evaporation and streamflow).

In summary, critical rate-limiting processes governing the local hydrologic cycle may be overlooked when applying methods in which the coupling at the water table is not specifically considered. This can have a large impact on the spatial estimates of ET, recharge and surface saturation.

There is great difficulty in evaluating models and assessing the importance of coupling because the relevant data is generally not available. The unusually detailed observations of Tóth (1966) provide one of few data sets useful for this purpose. In the following section we first describe the data and then present a model that will enable the issue of coupling to be explored.

12.2 METHODS

12.2.1 The Study Area and Available Data

The catchment chosen for this study (Trochu) is located in southern Alberta in the Canadian plains (Figure 12.1). This site was chosen because of the extensively documented fieldwork undertaken in the area by Tóth (1966), and because the poorly drained prairie topography emphasises the importance of three-dimensional groundwater circulation in determining the water balance (Levine and Salvucci, 1999a). The Trochu catchment is 16 km² in area with low relief (Figure 12.2). The maximum change in elevation from the water divide to the outlet is approximately 100 m and the maximum slope is approximately 6%. The general flow direction in the catchment area is west to east. The average slope from the farthest point in the catchment to the outlet is under 2%.

The soils in the study catchment are identified as thin black soils (combinations of silt, sand and gravel) developed on glacial drift material (Bowser et al., 1951). The bedrock consists primarily of nearly horizontal layers of sandstone and siltstone with some discontinuous layers of claystone and shale (Carlson, 1969; Tóth, 1966). The vegetation in the area is primarily cultivated rapeseed, alfalfa, and forage grasses with some small patches of aspen and willow trees along the ridge lines. Precipitation and temperature (New et al., 1999) both peak in the summer months (Figure 12.3), and the frost-free period typically begins in May and ends between September and October (Tóth, 1966). Average annual precipitation is approximately 440 mm.
Tóth (1966) studied this catchment extensively in an attempt to determine whether or not a correlation exists between physiographic features (e.g., vegetation types, presence of salt precipitates, moist depressions, well levels) and the direction of groundwater movement. In this work he provides a table of 152 field observations, 48 of which fall within the Trochu catchment. These observation points are labeled on Figure 12.2 (and the subsequent maps in this chapter) as R for recharge, D for discharge, C for creek bed and I for intermediate.

The categories were determined by applying Tóth’s (1966) criteria for evaluating surface observations and wells as follows. Sites classified as recharge through chemical analysis are those where ground or surface water testing showed low concentrations of dissolved minerals. Discharge sites were assumed where high concentrations of dissolved minerals were present. Observations of vegetation and surface salt deposition were used together to classify otherwise dry observation points as either discharge or recharge points (Figure 12.2). Where phreatic vegetation is present (e.g., slough grass, sedges and rushes)
Figure 12.1(b). Photograph of the Trochu catchment.

Figure 12.2. Surface elevation map of the 16 km² Trochu catchment with Tóth’s observation points. Observation points are labelled as follows: R = Recharge, D = Discharge, I = Intermediate, C = Creek bed. Elevation contours in metres. Grid labeling in metres.
without the presence of salt precipitates, the indication is that surface water recharges in this area. Phreatophytes with salt precipitates or salt precipitates alone indicate significant evaporation of groundwater, and so these points are classified as discharge points. Where two sets of criteria (i.e. water chemistry, vegetation, etc.) yielded different classifications, the point is labeled with both classifications.

Springs, seeps and flowing shotholes were all classified as discharge locations. Creek beds can be either gaining or losing reaches and so are not classified as either recharge or discharge. Piezometric classification was based on head to surface elevation comparisons. Where the head was near (<3 m) or above the surface elevation, a well was classified as being a discharge observation. Where head in a well was significantly lower than the surface elevation (>10 m), the piezometric determination was that the well was in a recharge zone. Between the two extremes wells were classified as being in intermediate zones. This criteria was not directly indicated in Tóth (1966), but rather inferred by comparing his reported measurements of depth to water against his final recharge–discharge map. Well head data was used as the sole determinant (i.e. without collocated chemical or botanical indicators) for approximately ten percent of the locations.

The resulting map (Figure 12.2) indicates a general pattern of recharge in the highlands (e.g. north-west and south-east borders of the basin and the north-west trending ridge in the northern third of the basin), and discharge in the low-lying areas, especially near the basin outlet. It is important to
remember that in this gentle prairie topography, mapped discharge locations include both surface contacts where liquid water seeps out of the ground and areas of persistent evapotranspiration of capillary rise. In the following section a model is described that accounts for the continuity and interdependence of such groundwater and vadose zone flows, and thus should be well-suited to reproduce the observed patterns.

12.2.2 Model Review

The position of the water table is dependent on the convergence (divergence) of groundwater flow, the amount of water being lost (gained) at the saturated/unsaturated interface, and the location of direct aquifer–surface water contacts (seeps, springs, lakes, etc.). The loss (gain) at the water table interface depends on the partitioning of fluxes in the vadose zone.

The vadose zone receives water from rainfall and capillary rise, and loses water through evapotranspiration and recharge. The rates of capillary rise, recharge, infiltration and evapotranspiration are all influenced by the soil moisture profile, depth to the water table, soil characteristics, and surface meteorological forcing.

Changes in vadose zone characteristics (soil moisture, matric potential, etc.) occur over short time and length scales, whereas the characteristic temporal and spatial scales for groundwater flow are generally larger. To ease the resulting computational burden of full saturated–unsaturated numerical simulations (e.g. Freeze, 1971; Paniconi and Wood, 1993), Salvucci and Entekhabi (1995) built on Eagleson’s (1978a–f) work to develop an equivalent steady state solution to the Richards equation bounded by a water table and driven by climate statistics (mean storm duration, intensity and frequency).

The climate statistics enter the model to parameterise the probability distributions of boundary conditions at the soil surface. Derived distribution techniques are used to average the resulting moisture fluxes over the storm–interstorm timescale, therefore providing an estimate of the time-averaged soil water flow through the vadose zone. This time-averaged flow, which may be downward recharge or upward capillary rise, forms a groundwater divergence boundary condition that is used to drive the spatially distributed groundwater model MODFLOW (McDonald and Harbaugh, 1996). As is discussed further below, the resulting distribution of water table depths influences the predicted vadose zone flow, and thus iteration is required to find the spatial distribution of water table depths for which the saturated flow divergence and vadose zone recharge are at equilibrium.

It is assumed in the model that interstorm evaporation and transpiration are driven by potential evaporation, but are influenced by both the mean root zone moisture content and by interstorm sources (capillary rise) and losses (recharge) below the root zone. The soil storage and infiltration capacities determine surface runoff through storage excess and infiltration excess mechanisms. The
storage capacity of the soil is determined by integrating the soil moisture deficit from the surface to the water table. The infiltration capacity is governed by gravity and matric potential gradients through a two-term Philip (1957) equation. The expected values of these soil and moisture dependent fluxes are derived subject to random precipitation events by integrating the event fluxes over probability distributions of storm intensity, duration and intermittency.

The fluxes through the vadose zone are determined by finding the equivalent steady state soil moisture profile that yields closure to the surface water budget. This profile is used as an initial condition for determining the infiltration and evapotranspiration capacity of the surface. It provides a coupling to groundwater by matching the time-averaged flux (recharge or discharge) and pressure head at the mean position of the water table. Salvucci and Entekhabi (1994a,b) show, by comparison with a long-term finite element simulation, that the equivalent steady state moisture profile solution closely approximates the long-term mean vadose zone flux over a wide range of soil texture and climate conditions.

An example solution, used below to drive MODFLOW over the Trochu catchment, is illustrated in Figure 12.4 (the climate and soil parameters used in this example are discussed later). Note that the dependence of the water budget

![Figure 12.4](image_url)

**Figure 12.4.** Simulated, long-term mean surface water fluxes for various depths to the water table. Silt-loam soil. $Z^* = 176$ cm. The first 45 centimeters above the water table are tension saturated. (From Levine and Salvucci, 1999a; reproduced with permission.)
on the depth to the water table ($Z_w$) occurs over a finite range from zero to the depth at which net recharge equals the maximum recharge rate. For this soil (silt–loam), the range of water table dependence is approximately 250 cm. There is a depth to the water table ($Z^*$) for each of the soil textures tested where the net recharge equals zero. At this depth, the long-term mean of transient groundwater losses due to capillary rise are balanced by the long-term mean of intermittent gains through recharge.

As the depth to groundwater decreases from $Z^*$ to the tension saturated zone (for the silt–loam soil between 176 cm and 45 cm), runoff increases, evapotranspiration increases up to the potential rate, and the net flux across the vadose zone–water table interface is upward. Runoff (solid line in Figure 12.4) increases due to a reduction in both the infiltration capacity and the storage capacity that occurs for higher initial soil moisture content and reduced depth to the saturated zone. Evapotranspiration (dotted line in Figure 12.4) increases to the climate limited rate because capillary rise from shallow water tables is large enough to replenish all water lost to evapotranspiration, even over long interstorm periods. As the depth to saturation increases from $Z^*$, evapotranspiration decreases, and recharge increases to balance the net of infiltration over the reduced evapotranspiration. The decrease in evapotranspiration occurs primarily because moisture supplied to the root zone by capillary rise decreases with increasing depth to water table.

Note that the water table dependence of vadose zone fluxes disappears for depth to water table greater than that at which both evaporation and net recharge reach their asymptotic values. This implies that water table–vadose zone coupling is insignificant in determining net recharge outside this range of depths.

The long-term mean of the net flux across the water table (the dashed line plotted in Figure 12.4) is used as an input to the groundwater flow model, MODFLOW, which is run in steady-state mode. This flux can be either positive for deep water tables (recharge) or negative for high water tables (discharge/capillary rise). This dependence provides a feedback between the surface water balance model and the groundwater flow model (Figure 12.5) whereby high water tables lose water to, and deep water tables gain water from, the vadose zone. The methodology for coupling the water table dependent recharge/discharge flux to the groundwater flow model is detailed in Levine and Salvucci (1999a).

12.2.3 Model Modifications

The vadose zone model presented in Salvucci and Entekhabi (1995, 1997) was modified to account for winter precipitation and snowmelt. Snowmelt is divided between storage excess runoff and infiltration according to the storage capacity of the vadose zone. Cold season evaporation is assumed negligible. The infiltration from snow melt is added to the flux to groundwater, assuming that its effect is mainly to increase soil moisture, and thus recharge, as a single pulse during a period of low evaporation.
The mean bare soil evaporation equation derived in Salvucci and Entekhabi (1995, 1997) has also been modified to account for vegetation. The interstorm transpiration is modelled as a two-stage process (unstressed and stressed), with the transition time dependent on soil type, soil moisture and rooting depth (Levine and Salvucci, 1999b). As for the other event-based fluxes in the model, the mean transpiration is calculated by integrating over the probability distributions of the time between storms. Because the modelled transpiration is more efficient than bare soil evaporation, warm season recharge is negligible and the source of deep-water recharge in Figure 12.4 is mainly melted winter precipitation. As will be seen in the coupled model runs below, lateral groundwater redistribution of this winter recharge makes up the moisture deficit (evaporation – rainfall – surface runoff) in shallow water table areas throughout the rest of the year.

Groundwater discharge at the surface is simulated by distributing drains over the surface using the drain package in MODFLOW (McDonald and Harbaugh, 1996). The stream networks predicted by the models were drawn using a flow accumulation algorithm with weights determined from the sum of groundwater discharge and surface runoff predicted at each cell.

Figure 12.5. Equilibrium conditions at discharging (a) and recharging (b) sections of a hillslope. The net recharge is positive (negative) for areas where the water table is deeper (shallower) than the depth at which mean annual groundwater flux is zero for the simulated soil and climate conditions. (From Levine and Salvucci, 1999a; reproduced with permission.)
12.2.4 Scenarios and Model Parameters

Following are short descriptions of estimated model parameters, discretisation choices, and soil–bedrock combinations chosen for model runs. Results, discussions and conclusions are documented in the subsequent sections.

The groundwater model was run with one soil layer and six bedrock layers at a horizontal grid spacing of 30 metres. The parameters in each layer were spatially uniform. The groundwater divide was assumed to coincide with the surface divide. The vertical spacing was variable, with the top layer thickness set to the depth of unconsolidated material (derived from Carlson, 1969) and the lower layers adjusted such that the modelled impermeable bottom was reached without any layer being more than 1.5 times as thick as the layer above it. The modelled impermeable aquifer bottom was placed at 500 m above sea level (an average depth of 385 m below the surface). Following Tóth (1966), the bedrock is approximated as a single homogeneous-isotropic unit. Drains were placed just below the surface in each column in order to simulate springs and seeps if the water table intersects the land surface at equilibrium conditions. All water leaving the saturated zone via the drains is assumed to exit the catchment as stream flow. A test run with 10 metre horizontal grid spacing yielded nearly identical results, most likely because the gentle topography of the prairie catchment is adequately described by 30 metre data.

The climate statistics required as input to the model (Table 12.1) were derived from the long-term record (29 years) for Lacombe, Canada (52°28′ N, 113°45′ W) provided by Environment Canada. This is the closest meteorological recording station with long-term precipitation and potential evaporation records. The potential evaporation data are pan evaporation multiplied by a single site-wide adjustment factor. All climatic variables are assumed to be spatially uniform. Results are presented for three simulations (Table 12.2) which include two soil types (silt-loam and clay-loam) over low conductivity bedrock and one soil type (silt-loam) over medium conductivity bedrock. The bedrock conductivities tested (Table 12.3) were chosen to cover the range of values estimated by Tóth (personal communication) in studies carried out over the Ghostpine and Three Hills Creek areas for the Edmonton and Paskapoo geologic formations. The soil types were chosen as representative of a silt-loam (similar to the local soil) and a clay-loam to demonstrate the effect that soil type has on surface/aquifer coupling. The

<table>
<thead>
<tr>
<th>Table 12.1. La Combe climate parameters used in the simulations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean time between storms*</td>
</tr>
<tr>
<td>Mean storm duration*</td>
</tr>
<tr>
<td>Mean storm intensity*</td>
</tr>
<tr>
<td>Mean evaporation*</td>
</tr>
<tr>
<td>Winter precipitation (snow water equivalent)</td>
</tr>
</tbody>
</table>

*Storm and evaporation statistics are for the 154-day average snow and frost-free period.

Source: Data from Environment Canada. Period of record is 1963 to 1992.
Brooks and Corey (1966) soil hydraulic parameters used to represent the soils are listed in Table 12.4. For simplicity the area is modelled with complete vegetation cover with an effective rooting depth of 45 cm. The model was also run for a two-dimensional cross-section in order to illustrate the variety of scales of circulation (local, intermediate and regional) making up the flow system. The two-dimensional results are presented in Levine and Salvucci (1999a).

### 12.3 RESULTS AND DISCUSSION

**12.3.1 Impact of Soil and Aquifer Hydraulic Parameters on Spatial Patterns Induced by Saturated Unsaturated Zone Coupling**

The influence of the coupling between vadose zone flux and depth to the water table is evident in the maps of simulated net recharge (Figures 12.6, 12.7 and 12.8). Case I (Figure 12.6) shows strong recharge (red) and discharge (blue) and many springs (white). Case II (Figure 12.7) shows strong recharge (red), discharge (blue), and springs (white) occurring over a smaller percentage of the area, and more extensive intermediate areas (yellow to pale orange) over which coupling strongly influences the height of the water table. Case III (Figure 12.8) is dominated by weak recharge (orange) and discharge (pale green) zones with extensive intermediate (yellow to pale orange) zones. The higher intensity of the recharge and discharge simulated in case I (and to a lesser extent case II) occurs because the higher conductivity allows greater flow through the aquifer, which in turn allows $Z_w$ to remain significantly below $Z^*$ over large areas.

<table>
<thead>
<tr>
<th>Case number</th>
<th>Soil type</th>
<th>Bedrock type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Case I</td>
<td>Silt-loam</td>
<td>Medium conductivity</td>
</tr>
<tr>
<td>Case II</td>
<td>Silt-loam</td>
<td>Low conductivity</td>
</tr>
<tr>
<td>Case III</td>
<td>Clay-loam</td>
<td>Low conductivity</td>
</tr>
</tbody>
</table>

**Table 12.3. Bedrock Conductivity**

<table>
<thead>
<tr>
<th>Bedrock type</th>
<th>Saturated conductivity (cm/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low conductivity</td>
<td>0.2</td>
</tr>
<tr>
<td>Medium conductivity</td>
<td>2.0</td>
</tr>
<tr>
<td>High conductivity</td>
<td>20</td>
</tr>
</tbody>
</table>
Reducing the soil conductivity and increasing the soil water retention ability (i.e. comparing case II, Figure 12.7 to case III, Figure 12.8) decreases the strength of the recharge (red to orange) and discharge (dark blue to pale blue and green) and yields larger midline areas (light orange and yellow). This compensation...

Table 12.4. Brooks and Corey parameters

<table>
<thead>
<tr>
<th>Soil Type</th>
<th>Saturated conductivity (cm/d)</th>
<th>Pore size distribution index</th>
<th>Effective porosity</th>
<th>Bubbling pressure head (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay-loam</td>
<td>3</td>
<td>0.44</td>
<td>0.45</td>
<td>-90</td>
</tr>
<tr>
<td>Silt-loam</td>
<td>30</td>
<td>1.2</td>
<td>0.35</td>
<td>-45</td>
</tr>
<tr>
<td>Sand-loam</td>
<td>300</td>
<td>3.3</td>
<td>0.25</td>
<td>-25</td>
</tr>
</tbody>
</table>

Source: Bras (1990)

Figure 12.6 Case I: Simulated equilibrium recharge, capillary rise, spring locations, and surface drainage. Colour bar scales from maximum recharge (red, +11 cm/year) to maximum capillary rise (blue, -46 cm/year). Seeps are denoted by white pixels. Observation points are labelled as follows: R = Recharge, D = Discharge, I = Intermediate, C = Creekbed. (From Levine and Salvucci, 1999a; reproduced with permission.)
occurs because the coupling effects for the clay-loam soil are active over a greater depth than for the silt-loam soil, requiring a greater depth to the water table to maintain recharge. The greater depth of interaction enables the feedback mechanism to regulate the water table height more strongly by restricting the saturated depth and hydraulic head in the relatively conductive soil layer, thus restricting the ability of the groundwater to drain away recharge. The reduced drainage potential, in turn, drives the water table back toward the equilibrium distance ($Z^*$) from the land surface.

Holding the soil type constant and increasing the bedrock conductivity (i.e. comparing case I, Figure 12.6 to case II, Figure 12.7) reduces the water table height, increases the strength and areal extent of recharge (red) and discharge (blue and green) zones, and shrinks intermediate zones (yellow to pale orange). The decrease in water table height under topographic highs weakens the impact that vadose zone–water table coupling has on the extent and strength of recharge. The decreased coupling under surface highs results in a larger amount of water
entering the flow system. This increased groundwater flow cannot be removed solely through evaporation in discharge areas, and thus surface seeps and springs develop where the water table is at the land surface. A good example of this is illustrated in case I (Figure 12.6) where springs (white) developed in most model predicted discharge areas, strong recharge (red) extends over large areas of the catchment, and only small zero net flux areas (pale orange) appear.

The significance of local variations in topography on the simulated spatial distribution of recharge and discharge areas can be seen in the more strongly coupled cases (II and III) (Figures 12.7 and 12.8). Note, for example, the discharge areas predicted at the base of steep slopes and in local convergence areas in the western portion of the study area. These areas are marked by a convergence of the local topography (Figure 12.2) and groundwater flow. The effect of the convergence of flow is to bring the water table closer to the surface, lowering the hydraulic gradient (slowing the rate of groundwater flux), and increasing the potential for discharge (through evaporation, springs or runoff production).

Figure 12.8. Case III: Simulated equilibrium recharge, capillary rise, spring locations, and surface drainage. Colour bar scales from maximum recharge (red, +11 cm/year) to maximum capillary rise (blue, −46 cm/year). Seeps are denoted by white pixels. Observation points are labelled as follows: R = Recharge, D = Discharge, I = Intermediate, C = Creek bed. (From Levine and Salvucci, 1999a; reproduced with permission.)
Case II (Figure 12.7) qualitatively captures many of the recharge and discharge points in the catchment. Well captured are the discharge points near the outlet of the catchment (eastern tip), the discharge areas along zones of groundwater convergence (west central portion of the catchment), and the recharge points near the higher elevation catchment boundaries (northern, south-western and southern boundaries). Poorly captured are the discharge points that occur along the side of the ridge forming the upper south-western boundary, the discharge areas that occur at the eastern edge of the ridge that extends south-eastward from the upland area at the northern boundary of the catchment, and the two recharge points that occur in the small valley that runs north from just west of the southern tip of the catchment (lower right corner).

Case III (Figure 12.8) captures the discharge areas that occur along the base of the ridge that runs south-east from the north-western boundary of the catchment to just above the centre of the catchment. These areas are missed by cases I (Figure 12.6) and II (Figure 12.7). This is an area where surface slope changes abruptly from a downhill pitch to more level ground (Figure 12.2). The water table in this region is fairly close to the surface for all the simulations, but for the silt-loam cases it is still below $Z^*$, and thus is predicted to receive water from the vadose zone (Figure 12.4). Only in case III is $Z^*$ great enough that the simulated mean flux at the saturated/unsaturated interface is toward the surface.

But case III fails to identify the recharge points located in the relatively flat upland area in the northern portion of the catchment (pale orange, Figure 12.8). This results from a competition between the relatively large depth to water table ($Z_w$) over which net recharge is negative for clay-loam soil ($Z^* = 370 \text{ cm}$) and the relatively shallow $Z_w$ (i.e. water table close to the ground surface) needed to provide the gradient and transmissivity for driving groundwater flow through the low conductivity aquifer. Together these effects act to restrict the strength and extent of recharge zones, as shown for simple hillslopes in Salvucci and Entekhabi (1995).

In case I (Figure 12.6), the higher lateral conductivity reduces the head gradient necessary to laterally drain groundwater recharge, resulting in a larger depth to the water table. This lower water table decreases the area over which vadose zone–water table coupling occurs. It results in larger and stronger recharge areas (red/orange), almost non-existent midline areas, very small (but strong) discharge areas (blue), and the development of significant areas of direct aquifer discharge to the surface (white). This case fails to capture the discharge areas at the transitions from steep hillsides to flat valley bottoms, but captures well the strong discharge along the centre of the lower valleys in the catchment (as does case II in general).

All of the model simulations miss the discharge point closest to the south-western boundary, approximately half-way up the catchment. This point is located midway down a hillside and may be the result of a discontinuity in the soil or aquifer material that creates a very localised flow system.

The matches between the model and field estimates for all three cases are summarised in Table 12.5. This table lists the percentage of observations for which the equilibrium model estimated the same direction of flow (recharge or
discharge) as was estimated from the field observations. As might be expected there is a tradeoff in which catchment parameters that lead to good estimates of recharge location underpredict areas of discharge and vice-versa. In part this results from the nature of the spatial structure; that is, recharge in one area of the basin impacts discharge in others. Note also that the only spatially distributed parameters in the model application were surface and bedrock topography. Calibration to individual field observations could be made by adjusting local soil hydraulic properties, but without independent measurements of these properties the significance of such calibration would be questionable.

**12.3.2 Stream Networks Generated by Equilibrium Model**

The model was also tested by comparing simulated to observed stream networks. The stream locations were predicted using a threshold-based accumulation algorithm of surface runoff and spring flow. The predicted stream network (shown in Figures 12.6, 12.7 and 12.8) thus reflects the surface and groundwater dynamics to a much greater degree than simple contributing area methods (e.g. algorithms that assume uniform runoff production across a basin). The stream locations predicted for case II (Figure 12.7) are nearby to Tóth’s (1966) three creek bed observation points and broadly correspond to the locations of streams plotted on the Canadian Centre for Mapping topographic map for the area (which, unfortunately, is at a scale where what constitutes a stream is somewhat subjective). In comparison with the mapped streams, the model overpredicts stream development in the area of groundwater convergence that runs north-south in the western portion of the catchment, and on the relatively flat areas in the southern third of the central area. Differences in stream locations generated from the simulated output may also be due to errors in the DEM, terrain analysis, or mapped locations.

Note also that the stream lengths and patterns change with the changes in patterns of recharge and discharge (Figures 12.6, 12.7, and 12.8). Case III (Figure 12.8) has the lowest overall groundwater flow, very little surface runoff, and high evaporation, which together result in the development of only a small stream network. The sources of the two streams that develop in this case are located in areas of strong groundwater convergence and high water tables and not in areas of weaker convergence. Case I (Figure 12.6) yielded a smaller stream network than case II (Figure 12.7) because the deeper water table of case I (resulting from

<table>
<thead>
<tr>
<th>Observation type</th>
<th>Case I</th>
<th>Case II</th>
<th>Case III</th>
</tr>
</thead>
<tbody>
<tr>
<td>R</td>
<td>100 %</td>
<td>86 %</td>
<td>57 %</td>
</tr>
<tr>
<td>D</td>
<td>67 %</td>
<td>78 %</td>
<td>89 %</td>
</tr>
</tbody>
</table>

**Table 12.5. Percentage of recharge and discharge points for which equilibrium model estimates match estimates based on field observations**
moderate bedrock conductivity) concentrates discharge in the lowest portions of the catchment. These results indicate that for this climate, high drainage densities would be expected in areas with highly conductive soils (to restrict evaporation) and low permeability bedrock.

Because of limitations in the mapped stream locations, the comparisons between mapped and predicted streams are of limited value. However, the strong sensitivity of the network shape and extent on soil and bedrock properties suggests that such comparisons could be useful for model calibration in situations where the streams are mapped, based on direct field observations.

12.3.3 Benefits and Drawbacks of Vadose Zone–Groundwater Coupling in Modelling the Spatial Patterns of Hydrologic Fluxes

In order to test whether water table coupling is a significant determinant in the spatial structure of recharge/discharge zones, or if the structure is constrained mainly by topography, a simulation was run holding the water table at the surface. As shown by Stoertz and Bradbury (1989) and others, holding the water table as a fixed boundary condition allows recharge to be estimated as the flow divergence at the surface of the aquifer. Note, however, that Stoertz and Bradbury (1989) proposed using the actual water table in such applications, while in our case the water table was held at the land surface (as in Ophori and Tóth (1989)). This was done because the water table data available from Tóth (1966) were too sparse.

Because holding the water table at the surface determines a spatial distribution of net recharge without consideration of vadose zone flow processes, comparing the skill of this method with the skill of the equilibrium model runs provides a simple test of the importance of two-way coupling. Furthermore, if it was found that the uncoupled method was able to represent the observed pattern of recharge and discharge, then topographic analysis alone could provide a useful estimate of spatial patterns of groundwater flow in a similar fashion to the way TOPMODEL (Beven and Kirkby, 1979) estimates spatial patterns in runoff (see Chapter 11).

The resulting patterns of recharge and discharge (Figure 12.9) do vary in general accordance with the observations of Tóth (1966). The amount of recharge necessary to maintain the water table can be much higher than the climate potentials. For example, recharge rates greater than 300 cm/yr are required to balance groundwater divergence in some locations (e.g. the deep red areas). Lowering the conductivity of the bedrock until the recharge areas have physically realistic intensities (i.e. less than annual precipitation) causes the disappearance of spatial patterns in recharge and discharge and eliminates almost all lateral flow (Figure 12.10).

Note that the fixed water table method also overemphasises the impact of small changes in topography by forcing the water table to reflect too closely the surface topography. This creates a more irregular recharge–discharge field, and
even highlights the triangular irregular networks formed in converting the surface elevation contour map to a digital elevation model.

As discussed in the introduction, it is not surprising that uncoupled groundwater models are susceptible to predicting recharge rates outside of reasonable bounds. On the other hand, coupled models such as that presented here suffer numerous limitations as well. Here the purpose was to show how observed patterns of recharge and discharge can inform modelling, but if the model were to be used for detailed predictions of spatially distributed catchment behaviour, one would need many more parameters in order to describe the meteorological forcing, vegetation, and unsaturated hydraulic properties. Other limitations (specific to the equilibrium model presented here) include the inability to simulate transient groundwater response and the inability to account for vertical soil heterogeneity in the vadose zone. This latter limitation could be addressed by replacing the Eagleson-type equilibrium water balance function (plotted in Figure 12.4) with the long-term mean flux partitioning predicted by a Richards-based soil–vegetation–atmosphere transfer model driven by long time

Figure 12.9. Simulated equilibrium recharge, capillary rise, spring locations, and surface drainage for an uncoupled model run with water table held at the surface. Soil and bedrock parameters as in Case II. Colour bar scales from maximum recharge (red, +316 cm/year) to maximum capillary rise (blue, −320 cm/year). Observation points are labelled as follows: R = Recharge, D = Discharge, I = Intermediate, C = Creekbed.
records of meteorological forcing. Such a hybrid approach would not be limited by soil homogeneity assumptions or other simplifying approximations that must be made to derive the analytical flux capacity relations in the Eagleson (1978a–f) and Salvucci and Entekhabi (1995) models. Whether or not relaxing these assumptions would lead to better model results, or simply more parameters to estimate, is an open question.

12.4 CONCLUSIONS

12.4.1 Modelling Spatial Patterns of Hydrologic Fluxes

The importance of accounting for two-way groundwater–vadose zone interaction when modelling the spatial distribution of surface fluxes has been demonstrated through a comparison against Tóth’s (1966) field observations and coupled and uncoupled model results. Allowing the position of the water table...
to influence both the surface water balance and the groundwater divergence field, and then constraining these two fluxes to balance over the long-term mean, defines an equilibrium condition for the catchment system. Past and current research in surface and groundwater hydrology (e.g. Tóth, 1963; Freeze and Witherspoon, 1966; Eagleson, 1978a–f; Stoertz and Bradbury, 1989; Salvucci and Entekhabi, 1994a,b, 1995; Kim et al., 1999) has demonstrated and/or supposed that this equilibrium state forms an estimate of long-term mean conditions. Under this coupled and time-averaged condition, water balance partitioning in any one part of a basin can influence the partitioning at distant points. This behaviour imparts strong diagnostic value to spatially distributed field observations.

12.4.2 Comparison with Observations

The simulated recharge and discharge patterns match field observations of both recharge and discharge best for case II, but not as well for the more and less permeable conditions of cases I and III. Case I had more permeable bedrock and allowed the groundwater to reside deeper in the ground. This minimised saturated–unsaturated zone coupling and resulted in better prediction of recharge areas but underprediction of discharge areas. Case III had less permeable bedrock and more retentive (clayey) soils. The reduced permeability forced the water table closer to the surface (to increase hydraulic gradients and transmissivity), and thus brought more of the catchment area under saturated–unsaturated coupling. As a result, this case underpredicted recharge areas but captured most observed discharge areas.

Together the results indicate that coupling is an important factor in determining the spatial structure of recharge/discharge zones. Model estimates of the location of recharge and discharge with the water table shape determined by topography alone (i.e. without two-way interaction) also matched the field observations reasonably well, but appeared, in a qualitative sense, unrealistically heterogeneous and could be achieved only with physically unrealistic recharge intensities. In contrast, the fluxes predicted by the coupled model are within the bounds imposed by the climate forcing (i.e. model predicted recharge is less than rainfall, and model predicted discharge, except in cases of surface water contact, is less than potential evaporation). This condition is not met by the uncoupled model results where the water table was held fixed at the surface.

12.4.3 Impacts of Saturated–Unsaturated Zone Coupling on Catchment Modelling

Conclusions specific to the modelling experiments, which may be transferable to other catchments, can be summarised as follows:
The overall effect of the vadose zone feedback mechanism (at a point) is to increase recharge if the depth to the water table is greater than equilibrium depth \( Z^* \) and to increase capillary rise if it is less than \( Z^* \).

The spatial patterns of recharge/discharge are dependent on both the aquifer permeability and the depth (of order \( Z^* \)) over which strong surface coupling exists. The patterns are more dependent on the local surface topography when strong coupling exists (Figures 12.7 and 12.8), and on the regional topography when weaker coupling exists (Figure 12.6).

Low bedrock conductivity limits lateral redistribution of water and requires the water table to rise under topographic highs in order to drain recharging water. Higher water tables, however, restrict net recharge through both enhanced capillary rise to the root zone and increased runoff generation. Thus limited transmissivity increases the horizontal extent over which coupling at the water table plays a role in keeping the flow system in balance.

This work has shown that an equilibrium, coupled groundwater–vadose zone model is able to represent the long-term spatial patterns of recharge and discharge in the prairie landscape of the Trochu catchment. This conclusion was made possible by the extraordinarily detailed field data of Tóth (1966) that was unusual not only in its spatial detail, but also in the nature of the measurements. These were generally integrators of long-term response, such as vegetation and soil or water chemistry, and so were ideal for the testing of the equilibrium-style model used in this study. Indeed, without data of this type, testing the spatially distributed predictions from equilibrium models would be impossible.

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