1.10 THE OVERVIEW OF ATMOSPHERIC AND SURFACE WATER COUPLING TO REGIONAL GROUNDWATER MODELS IN SEMI-ARID BASINS

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1. INTRODUCTION

This paper describes a series of data collection activities undertaken at the Lewis Springs site on the San Pedro River in Southeastern Arizona, USA. The data collected at the site will be analyzed and used to modify the groundwater flow modeling modules for stream/aquifer interaction and evapotranspiration. The Lewis Springs Site, the first of a series of sites on the San Pedro River, is representative of groundwater system with a gaining stream. At present, research at the site is in the data collection phase and only limited preliminary analysis has been undertaken (Goodrich et al (a), this issue).

2. PRESENT MODULES

In arid and semi-arid regions, groundwater and surface water interactions maybe intricately coupled with evapotranspiration processes in narrow bands of vegetation along streams. These vegetative bands are referred to as riparian corridors, and the stream systems with which they coexist are usually perennial or intermittent, but some may even be ephemeral.

In the present state of groundwater modeling, the stream and evapotranspiration within these riparian corridors are modeled as source terms (they are actually boundary conditions that have been converted to source terms). The source terms are treated as head dependent in a piecewise linear fashion.

It is assumed that the flow, Q_R, between an aquifer and a stream is governed by Equation 1,

$$Q_R = \begin{cases} C_R(H_R - h_A) & h_A > H_B \\ C_R(H_R - H_B) & h_A \le H_B \end{cases} \tag{1}$$

where C_R is the river bed conductance [L^2/t], H_R is the river stage [L], h_A is the head in the aquifer, and H_B is the elevation of the bottom of the river bed [L](see Figure 1).

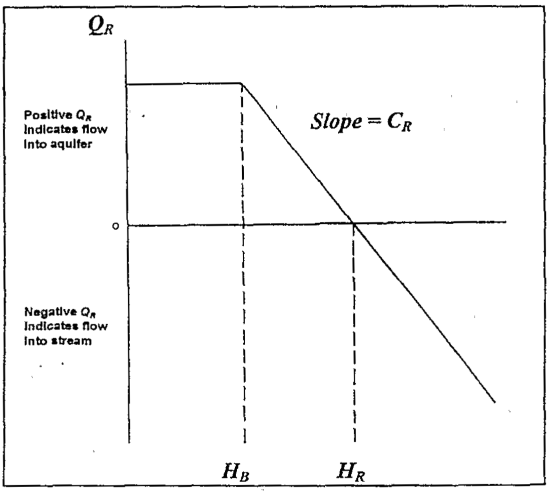

Figure 1. Leakage, Q_R, through streambed into aquifer, after McDonald and Harbaugh, 1988

It is assumed that the evapotranspiration loss, Q_{et}, is governed by Equation 2,

$$Q_{et} = \begin{cases} Q_{\max} & h_A > H_S \\ 0 & h_A < H_S - d \\ Q_{\max}\dfrac{h_A - (H_S - d)}{d} & (H_S - d) \le h_A \le H_S \end{cases} \tag{2}$$

where Q_{max} is a maximum evapotranspiration rate, H_s is the elevation of a maximum evapotranspiration surface above which evapotranspiration is assumed to be at a maximum value and constant, and d is an extinction depth

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below which no evapotranspiration is assumed to occur (see Figure 2). Because the finite-difference grid size in a groundwater flow model can be quite large (sometimes in square miles), the Q_{max} is an area-averaged aggregate over all the plant species within the gridded area.

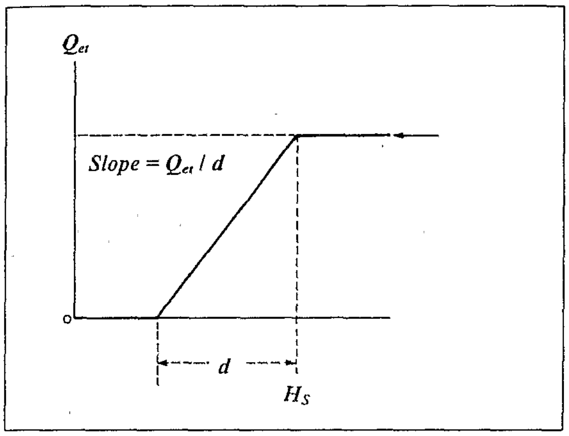

Figure 2. Aggregated evapotranspiration, Q_{et}, after McDonald and Harbaugh, 1988

The stream and the evapotranspiration source terms are introduced into the groundwater flow model as separate modules and there is no explicit coupling between the two terms. All coupling is thus an artifact of coexistence of the modules within the model, and the fact that both source terms are aquifer-head dependent.

3. TIME SCALE PROBLEM

The interrelation of groundwater, surface water and evapotranspiration creates a time scale problem. Groundwater modelers assume that streamflow is composed of two time-scale processes: a rapid time-scale process, the runoff; and a slow time-scale process, the baseflow. The baseflow time scale is equivalent to the groundwater time scale and may be measured in periods of months or years. The fast time scales are measured in hours or days. In many groundwater flow models, it is assumed that the fast time-scale runoff does not interact with the groundwater system, the runoff can be ignored, and only the baseflow component interacts with the groundwater system.

In arid and semi-arid basins that contain unregulated streams, an indication of the time-scale difference is provided by the comparison of mean-annual streamflow with median-annual streamflow. Several orders of magnitude can exist between the two statistical flows. It is not uncommon to find a stream with 100 cfs mean flow and 1 cfs median flow. Furthermore, examination of the flow duration curve for the stream is likely to show that the mean-flow event (or greater) occurs only a very small percentage of the time (Figure 3).

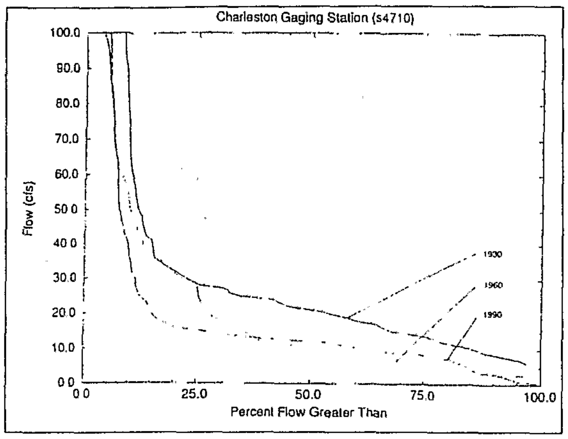

Figure 3. Flow duration curve for years 1930, 1960 and 1990 for the Charleston gauge on the San Pedro River

Baseflow calculation is an arduous task requiring the averaging of the separation compilation of many streamflow events over many years. Statistical surrogates are sometimes used in lieu of baseflow; for example, in many western streams the annual averaged 7-day low flow is an excellent candidate.

Evapotranspiration processes are estimated using a combination analysis that consist of examining aerial photographs to determine plant species density and distribution, applying a technique such as the Blaney-Criddle method (1950) to determine the evapotranspiration for the species, and then weighting the evapotranspiration by the densities and distributions to obtain an areal average.

When the groundwater flow model is run with baseflow as the streamflow and with evapotranspiration as estimated by the above procedure, the simulation can produce excessive streamflow loss predictions. Thus, the model predicted stream or baseflows are less than the calculated baseflows from the streamflow data. The difference between simulated and calculated streamflow is due to the fact that a portion of the evapotranspiration is driven by the runoff, which as been excluded from the model. Thus, if one creates a dichotomy of the streamflow into baseflow and runoff, then likewise, one must create a dichotomy of the evapotranspiration between water taken from the water table that interacts with stream baseflow, and water taken from the unsaturated zone that is commonly driven by both precipitation at the site and the runoff.

4. LEWIS SPRING SITE

At the Lewis Springs site, a set of experiments were designed to examine the evapotranspiration processes. The experiments are to determine how much of the plant transpiration comes from the vadose zone and how much comes from the water table. Williams et al (this issue) determine which one is the primary source of water by isotopic techniques. Because the evapotranspiration processes vary seasonally, it is necessary to determine seasonal variations. Current groundwater models do not typically treat seasonal variations. Diurnal variations described by Williams et al (this issue), furthermore, are influenced by radiation inputs, humidity, and possible feedback effects on stomatal conductance.

The Lewis Springs Site experiments determine soil moisture seasonal variations in the vadose zone, seasonal water table variation in the saturated zone, seasonal stage and discharge variations in the stream, and seasonal variations in evapotranspiration processes.

Seasonal variations were determined by 5 synoptic runs (32 to 48 hour periods of intensive data collection) that occurred in March, April, June, August and October. The March synoptic run was a shakedown run. The April run was prior to leaf out. The June run was prior to the monsoon and represented the time period with the highest temperatures and lowest humidity and thus the highest evaporative demand. The August run was during the monsoon and encompassed periods of runoff and the October run was post monsoon and was selected to represent end of the growing season conditions. The April, June, August and October synoptic runs were coordinated with isopotic and sapflow measurements and remotely sensed to estimate large area estimates of riparian evapotranspiration over the 60 kilometer San Pedro riparian corridor (Williams et al, Moran et al, Qi et al, and Hipps et al, this issue).

4.1 Vadose zone Measurements

The near-stream (within the cottonwood-willow forest gallery) seasonal and spatial variations of soil moisture in the vadose were estimated using tensiometers (24), water content reflectometry probes (20) which were continuously recorded, and neutron probe access tubes (12) distributed as shown in Figure 4, were measured periodically. The near-stream soil moisture measurement devices were located in the cottonwood-willow gallery (see Whitaker et al, this issue).

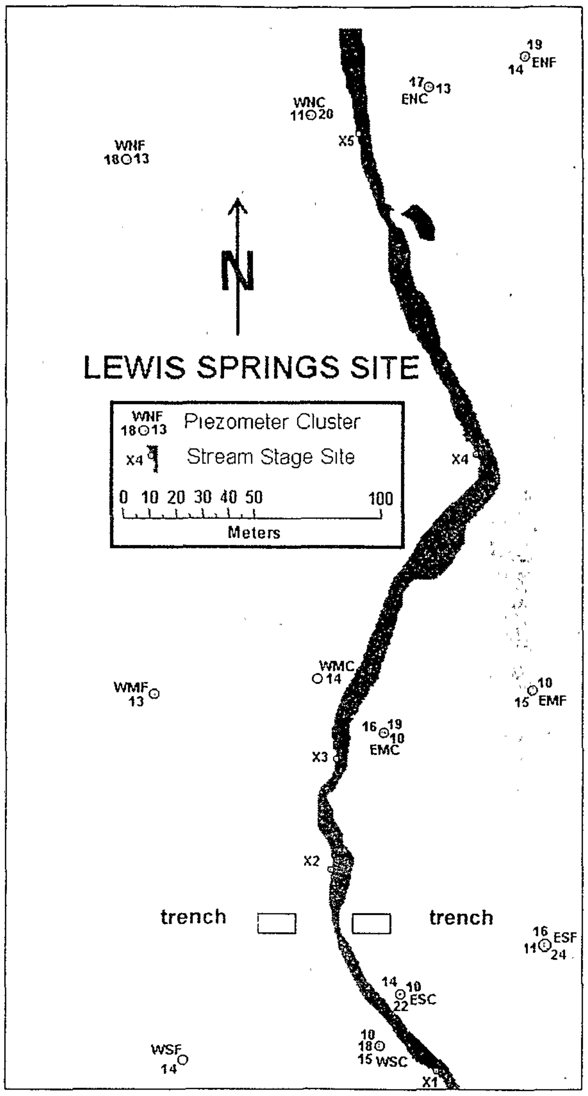

Figure 4. Lewis Springs Site map.

The far-stream (the sacaton and mesquite vegetation complexes outside the cottonwood-willow gallery but inside the historic floodplain) seasonal and spatial variations of soil moisture in the vadose were estimated using water content reflectometry probes (5) buried in two trenches and are distributed as shown in Figure 4 (see Whitaker et al, Moran et al and Hymer et al, this issue). The far-stream soil moisture measurement devices were located in a sacaton grass and mesquite shrub area (see Scott et al, Moran et al and Hymer et al, this issue).

4.2 Water table measurements

Seasonal and spatial distribution of the water table were estimated using piezometer nests (12 with 1 to 3 piezometers per nest for a total of 31 piezometers) and wells (6) distributed as shown in figure 4. Because the water table is located in a shallow alluvial aquifer that interacts with a deeper regional aquifer, some piezometers and wells were also located in the upper portion of the regional aquifer (see Mac Nish et al, this issue). Some of the piezometer nests were located in the sacaton-mesquite area. During the synoptic periods, water levels were measured hourly on all piezometers, and were continuously recorded on the 6 wells (Mac Nish et al, this issue).

4.3 Stream flow measurements

Stream stage was determined using staff gages in March, and stilling tubes in the later synoptics. Streamflows were measured periodically by pygmy meter in all the synoptics, continuously in the June synoptic with a flume, and continuously in all synoptics using a constant rate dye injection technique. During the synoptics, stages were recorder hourly, and samples were taken hourly at five points in the study reach for dye concentration analysis. During the June synoptic run a small H-flume was installed at the upstream end of the main study reach (Mac Nish et al, this issue).

4.4 Evapotranspiration measurements

Evapotranspiration in the near-stream cottonwood-willow galley was measured using sapflow meters (see Snyder et al, this issue) and LIDAR (see Cooper et al, this issue).

Evapotranspiration in the far stream sacaton-mesquite area was measured using two Energy Budget-Bowen Ratio systems both mounted on towers, one above an extensive area of sacaton grass and one above a stand of Mesquite bosque. Theses two systems and an Automatic Weather Station were used to provide near-continuous measurements of evapotranspiration and near-surface weather variable for a complete annual cycle (see Scott et al, this issue)

Evapotranspiration may also measured indirectly by analyzing the diurnal fluctuations of streamflow during the synoptic runs.

5. FUTURE RESEARCH

From these coordinated studies, the following results for modeling ground and surface water systems are anticipated:

1) better estimates of streambed conductance for riparian systems (see C_R, Equation 1),
2) a more accurate curve configuration for surface/ground water interactions (see Equation 1),
3) seasonal parameter estimates for both surface and groundwater systems,
4) dichotomy for the evapotranspiration processes between the vadose zone and the water table (see Equation 2)
5) seasonal evapotranspiration curves for water table extractions (see Equation 2).

The Lewis Spring Site is representative of a gaining stream site. Further sites are needed, particularly in a losing stream region, to insure proper under understanding of a highly complex physical system. To the groundwater modeler, the Lewis Spring Site is representative of a single grid cell, whereas the model is composed of a multitude of cells that interact with the surface water system.

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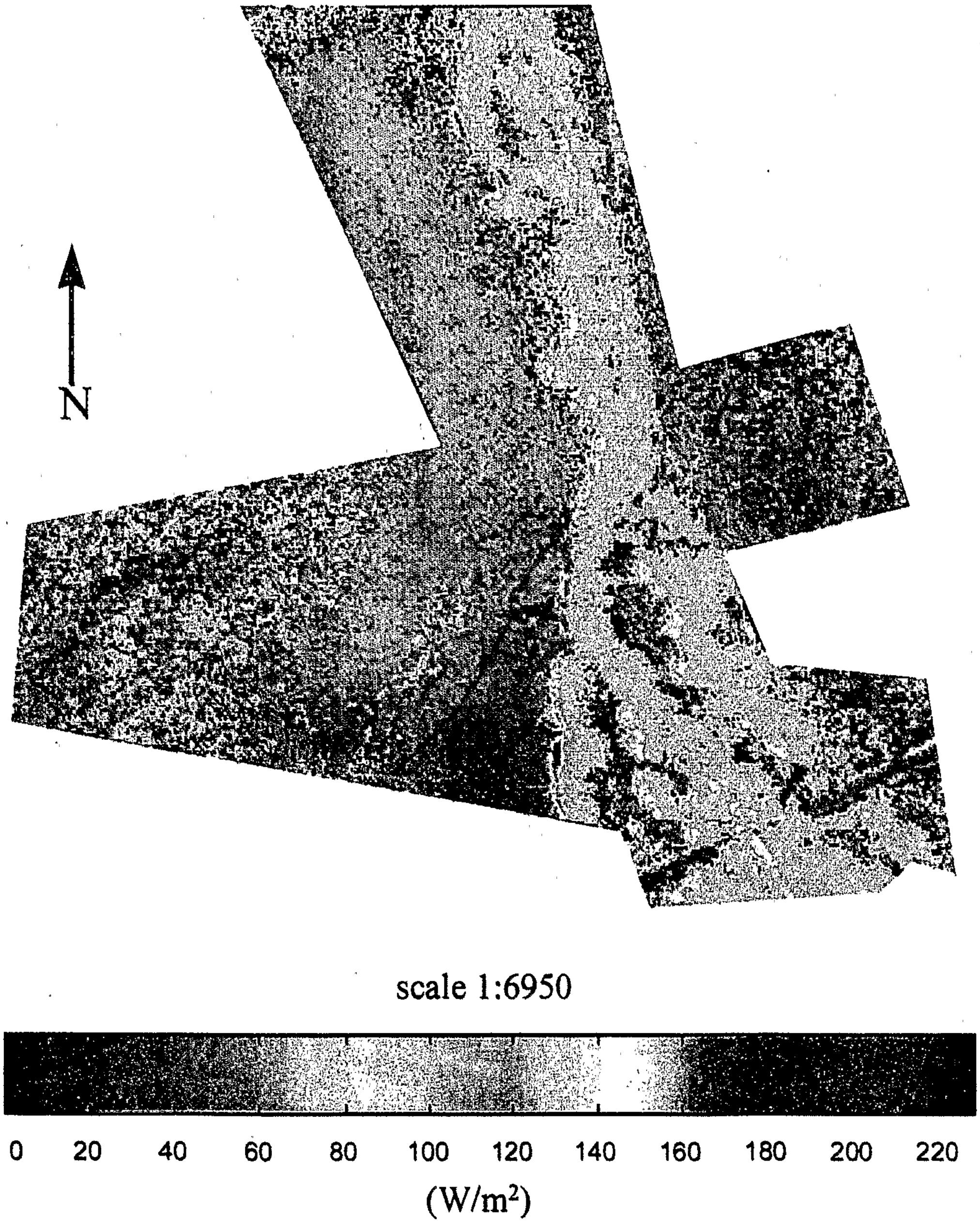

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