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LAKE AND GROUNDWATER PALEOHYDROLOGY:
Use of Groundwater Flow Theory to Explain Past Lake Levels in West-Central Minnesota

by: James E. Almendinger

minnesota water resources research center

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ABSTRACT

Investigation of a simple analytic model of an interfluvial water table demonstrates that a shift in groundwater recharge N changes the water table elevation the most near the middle of the interfluve. Consequently, lakes lying farthest from rivers are most vulnerable to lake-level change. The partial derivative of groundwater head with respect to N, the "positional sensitivity," is quantified for the simple model as a function of position across the interfluve. Despite its simplicity, the positional sensitivity of the model has some predictive value for water-table and lake-level changes in a sandplain in west-central Minnesota.

Lake levels are also a function of surficial hydrology. "Lake pumping" is symbolized by γ and defined as the net removal of water from a lake by hydrologic processes acting at the lake surface, namely evaporation minus direct precipitation and minus any input from overland runoff that reaches the lake. Investigation of a simple analytic groundwater model of a circular lake next to an infinitely long river shows that the sensitivity of the lake level to a change in γ is proportional to the radius of the lake and its distance from the river. The analysis also indicates that lakes lying in highly permeable substrates are not very sensitive to changes in γ.

The response of a lake level to a shift in climate depends on characteristics of surficial and groundwater hydrology that are unique to that lake. Determination of the past levels of several lakes, rather than just one, should help provide a more nearly unique reconstruction of past hydrology and climate. Analysis of the sediments of several closed-basin lakes lying in the Parkers Prairie sandplain in west-central Minnesota indicates that lake levels were lowest about 8.5 to 8 ka. I manipulate the N and γ of a steady-state analytic-element groundwater model such that the modeled water table coincides with the paleo-lake levels for a given past time. Model results indicate that lake levels at 8.5 to 8 ka can be explained primarily by reducing N to 40% of the modern value, coupled with a γ of about 20 to 30 cm yr^{-1}. By 6 ka N had increased to 50 to 80% of the modern value allowing most lakes to rise in level, but γ may also have increased forcing at least one lake to remain nearly dry.

CHAPTER 1

GROUNDWATER CONTROL OF CLOSED-BASIN LAKE LEVELS: 1. EFFECT OF DIFFERENT AQUIFER RECHARGE RATES UNDER STEADY-STATE CONDITIONS

ABSTRACT

Investigation of a simple analytic model of an interfluvial water table demonstrates that a shift in groundwater recharge changes the water table elevation the most near the middle of the interfluve. The partial derivative of groundwater head with respect to recharge, the "positional sensitivity," is quantified for the simple model as a function of position across the interfluve. Despite its simplicity, the positional sensitivity of the model has some predictive value for water-table and lake-level changes in a sandplain in west-central Minnesota. Knowledge that lake-level sensitivity can be directly proportional to the distance from the lake to the nearest river may help identify which lakes on the landscape are vulnerable to lake-level change and may help in the interpretation of paleo-lake levels.

INTRODUCTION

Aside from seasonal fluctuations in level of usually a meter or less, lakes are often considered to be rather stable features of the landscape. Lakes with outlets are especially stable, while closed-basin lakes (those without outlets) exhibit more fluctuations in level. Rivers might be perceived as being unstable in level, given the floods that can result from extraordinary events of snowmelt and precipitation. However, when viewed over longer time periods (decades to millennia) under steady-state hydrologic conditions, rivers tend to be the more permanent and stable hydrologic features, assuming fluvial downcutting and deposition are minimal. In contrast, over the long term the water table and the closed-basin lakes connected with it may change many meters in elevation.

While closed-basin lakes can exist under many circumstances, one of the most common settings is in fairly permeable substrates under a climate in which evaporation more or less balances precipitation. The upper Midwest contains many such closed-basin lakes, and slight shifts in the regional climate on a time scale of decades or even less can change the steady-state elevation of these lakes by several meters. For example, Big Marine Lake in eastern Minnesota rose over 3 m in about 40 years (Brown, 1985); School Section Lake in central Minnesota rose over 2 m in less than five years (Mohring, 1986); and Devils Lake in eastern North Dakota fell almost 12 m during the period 1867 to 1941, and then rose about 4 m from 1941 to 1951 (Swenson and Colby, 1955; the lake has continued to rise since their report). Such lake-level changes can cause extensive property damage or devaluation. It would be valuable to identify those lakes in the landscape that are most vulnerable to such changes to allow closer monitoring and perhaps control.

Because of the obvious, although complicated, relationship between modern lake levels and short-range climate, past lake levels have been studied in an effort to infer past climate. Most of these efforts have focused on the surficial hydrology of the lake's watershed by using a water-budget equation or a water-and-energy-budget equation to find the possible climatic conditions that could have caused the past lake levels as indicated by the geologic evidence (*e.g.*, Kutzbach, 1980; Benson, 1981; Roberts, 1983; Street-Perrott and Roberts, 1983; Street-Perrott and Harrison, 1985; Winkler *et al.* , 1986). These studies are limited to situations where the shifts in the water table are simple enough so that groundwater flow need not be modeled, namely situations in which groundwater divides are congruent with surficial watershed boundaries. However, a great many lakes are connected with a regional interfluvial water table, and the lake levels change in direct response to shifts in the elevation of the water table. In such cases, the effect of changing water-table elevations must be factored out, perhaps through the use of a groundwater model, before a surficial water-budget equation could be applied with validity. Thus paleoclimatic inferences would benefit greatly from knowledge of how lake levels were sensitive to changes in the water table.

This paper demonstrates how a simple interfluvial water table is sensitive to changes in recharge. To this end, I examine a steady-state analytic model (a simple equation, really) of perhaps the simplest possible interfluvial configuration: a strip of surficial aquifer bounded on two sides by infinitely long canals of constant head. Not surprisingly, the results show that changes in recharge affected the water table the most near the middle of the interfluve. Despite the simplicity of the model, I apply the results to an actual field situation in west-central Minnesota, and I explore the limits of this applicability. Throughout the paper I assume that lakes are small (a few to roughly 100 ha in area) and that they are merely indicators of the water-table elevation, much like large-diameter monitoring wells. The complexities of lake-groundwater interactions are being investigated with increasing detail by others (McBride and Pfannkuch, 1975; Winter, 1976, 1978; Pfannkuch and Winter, 1984; Winter and Pfannkuch, 1984), and must be considered when one deals with the detailed hydrology of individual lakes.

METHOD: SIMPLE-SYSTEM EQUATION FOR GROUNDWATER POTENTIAL

The actual sensitivity of the piezometric head (ϕ) to a change in recharge (N) may be defined as $\partial\phi/\partial N$ (see Table 1.1 for a summary of symbols used). To evaluate this expression, a system of very simple geometry must be defined so that the equation describing head as a function of position will be short and tractable. The system I chose imposes uniform recharge over a strip of surficial aquifer bounded on two sides by infinitely long parallel rivers (canals, really, of constant head along their lengths); all groundwater flow in such a system is linear and perpendicular to the rivers (Fig. 1.1). The rivers need not have the same head. If groundwater potential in a water-table aquifer is defined as $\Phi = 0.5\ k\ \phi^2$ then

Table 1.1. Definitions of parameter symbols and values of constants

Symbol	Definition or value (unit of measurement)
ϕ	= water-table head (m)
ϕ_A	= head at river A = 30 m
ϕ_B	= head at river B = 20 m
k	= hydraulic conductivity (m s^{-1})
k_1	= 1E-3 m s^{-1}, coarse sand and gravel
k_2	= 0.5 k_1 = 5E-4 m s^{-1}
Φ	= 0.5 k ϕ^2 = groundwater potential (m^3 s^{-1})
Φ_A	= 0.5 k $\phi_A{}^2$ = groundwater potential at river A (m^3 s^{-1})
Φ_B	= 0.5 k $\phi_B{}^2$ = groundwater potential at river B (m^3 s^{-1})
N	= groundwater recharge (m s^{-1}, or cm yr^{-1})
N_1	= 5E-9 m s^{-1}, which ≈ 15.8 cm yr^{-1}
N_2	= 2 N_1 = 1E-8 m s^{-1}
D_A	= distance from river A (m)
D_B	= distance from river B (m)
L	= $0.5(D_A + D_B)$ = half the distance from river A to river B (m)
D_{AL}	= D_A/L = distance from river A relative to L (unitless)
S_N	= $(\partial\Phi/\partial N)/L^2 = (k\ \phi\ \partial\phi/\partial N)/L^2$ = dimensionless positional sensitivity of the groundwater potential with respect to a change in recharge (unitless)
$\partial\phi/\partial N$	= $\{(C\ L^2)/(k\ \phi)\}\ S_N$ = actual positional sensitivity of the water table with respect to a change in recharge (m/(cm yr^{-1}), or m/(m s^{-1}) if C is omitted)
C	= 3.171E-10 (m s^{-1})/(cm yr^{-1}), merely a conversion factor
N_o	= 5E-9 m s^{-1} ≈ 15.8 cm yr^{-1}, estimate of modern-day recharge used in the full groundwater model of the Parkers Prairie sandplain
k_o	= 2.25E-3 m s^{-1}, estimate of hydraulic conductivity used in the full groundwater model of the Parkers Prairie sandplain

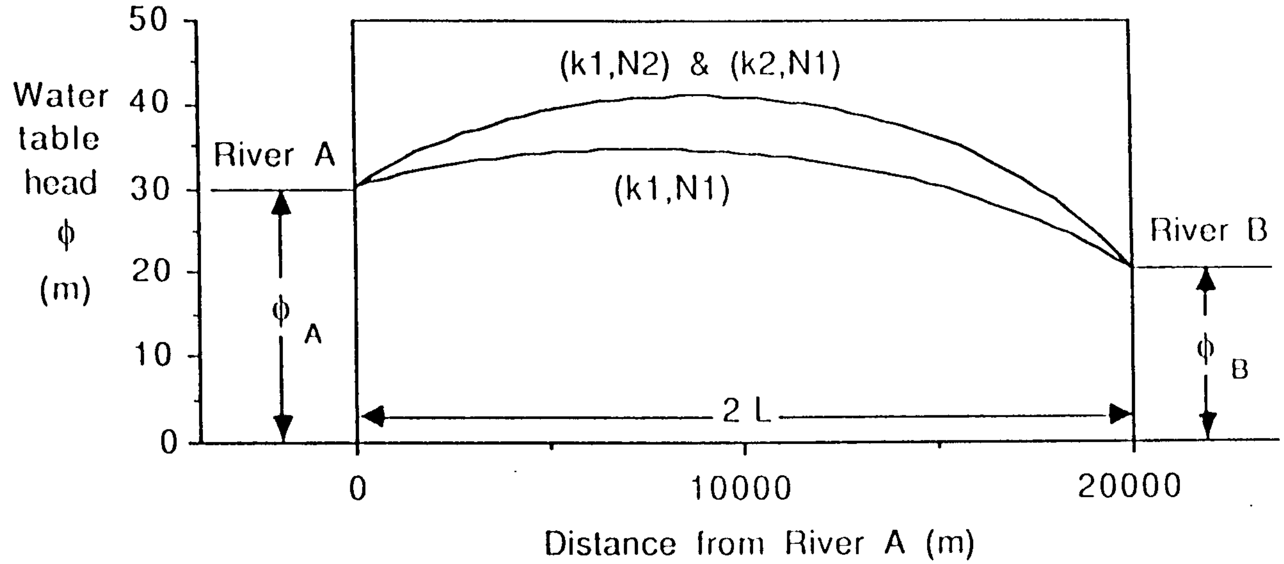

Fig. 1.1 Plots of equation 1.1: examples of several water-table configurations in a cross-section of a strip of aquifer that receives uniform recharge and is bounded by infinitely long parallel rivers. N1 < N2, and k1 > k2; meanings of symbols and values of parameters are given in Table 1.1.

$$\Phi = (-N/2)(D_A^2 - 2D_AL) - (\Phi_A - \Phi_B)(D_A - L)/2L + (\Phi_A + \Phi_B)/2 \qquad (1.1)$$

describes the groundwater potential as a function of position and recharge in the strip of aquifer, given the heads of the two rivers and the hydraulic conductivity of the aquifer (see Table 1.1 for meaning of symbols used; see Strack, 1988, for a derivation of essentially this same equation). This equation models steady-state, two-dimensional flow in the horizontal plane and is restricted by the typical assumptions of isotropic and uniform hydraulic conductivity, uniform recharge, and the Dupuit-Forchheimer relation ($\partial\phi/\partial z = 0$, and no resistance to flow in the vertical direction). The elevation of the water table is proportional to the quantity N/k, and thus the height of the interfluvial water-table mound may be increased either by increasing N or by decreasing k (Fig. 1.1).

RESULTS: POSITIONAL SENSITIVITY OF THE WATER TABLE

Dimensionless positional sensitivity

To find the sensitivity of the groundwater potential to recharge in the simple system depicted in Fig. 1.1, equation 1.1 may be differentiated with respect to recharge N. To make the result independent of the dimensions of the system, all lengths are measured in units of L (*i.e.*, all lengths are divided by L), and the resulting partial derivative is:

$$S_N = (\partial\Phi/\partial N)/L^2 = (k\phi\partial\phi/\partial N)/L^2 = -0.5(D_{AL}^2 - 2D_{AL}) \qquad (1.2)$$

in which $D_{AL} = D_A/L$ (see Table 1.1). S_N is the "dimensionless positional sensitivity" of the groundwater potential with respect to recharge; *i.e.*, S_N represents a sensitivity that is a function *only* of position in the watershed, and that position is independent of (= normalized by) the dimensions of the watershed. Qualitatively, S_N behaves as expected: sensitivity is zero at the rivers because the water table is forced to maintain a constant head there, and sensitivity is maximum farthest away from the rivers at the exact middle of the interfluve.

Actual positional sensitivity

The "actual positional sensitivity" of the water table to a change in recharge is the quantity $\partial\phi/\partial N$, which may be obtained by rearranging equation 1.2:

$$\partial\phi/\partial N = \{(C\,L^2)/(k\,\phi)\}\,S_N \qquad (1.3)$$

(C = 3.171E-10 m s^{-1}/cm yr^{-1}, and is just a factor that allows N to be measured in the simpler units of cm yr^{-1} rather than m s^{-1}; omit C if N is in units of m s^{-1}.) To solve equation 1.3, geologic and hydrologic information about the aquifer will have to be known or assumed. Namely, for the aquifer L and k must be approximated in some way; and for the point in question, ϕ must be known or evaluated from equation 1.1, and S_N must be evaluated from equation 1.2 (or by reference to Fig. 1.2). In the plot of equation 1.3 (Fig. 1.3) I used the same values for L, k, and N as in Fig. 1.1.

The shape of $\partial\phi/\partial N$ as a function of position (Fig. 1.3) is very similar to that of S_N, with one slight modification. As shown in equation 1.3, $\partial\phi/\partial N$ is inversely

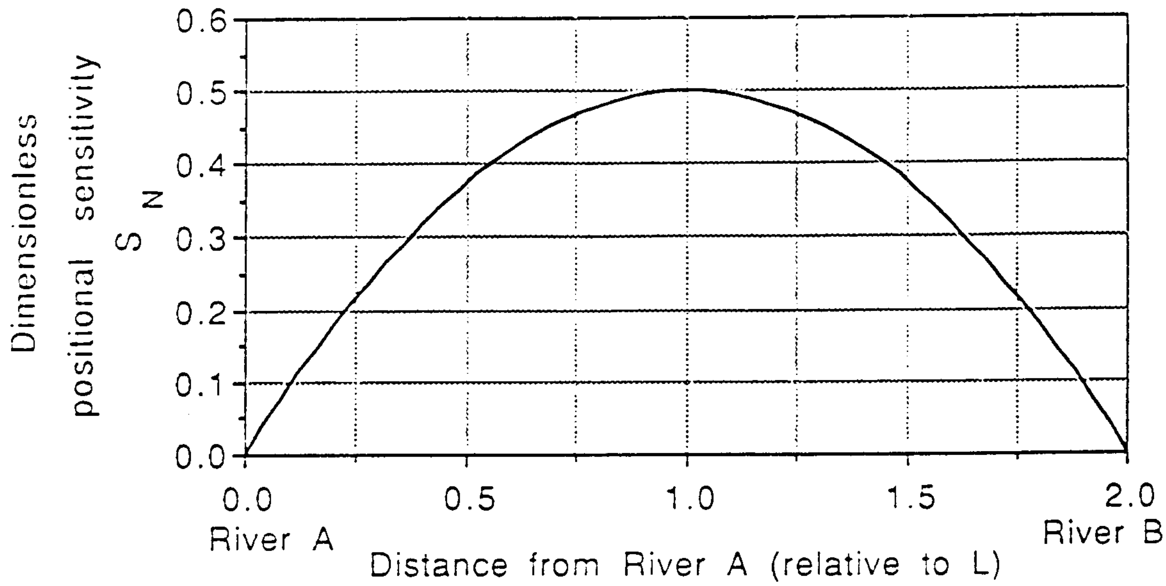

Fig. 1.2. Plot of equation 1.2: the sensitivity, or vulnerability, of the groundwater potential to a shift in recharge, as a function of position (distance from river A) in the simple aquifer modeled by equation 1.1 and depicted in Fig. 1.1. Both axes are dimensionless, because all linear units were measured in terms of L = half the distance from river A to river B (see Fig. 1.1 and Table 1.1).

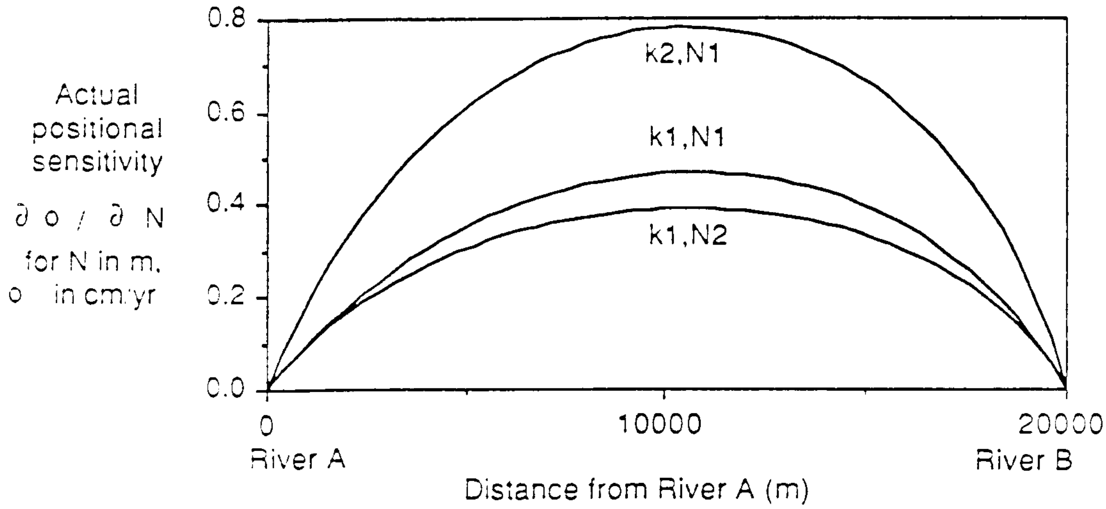

Fig. 1.3. Plots of equation 1.3: examples of the actual sensitivity of the water-table level to a shift in recharge, as a function of position (distance from river A) in the simple aquifer as modeled by equation 1.1. These three plots correspond to the same three examples depicted in Fig. 1.1; actual values for the parameters are given in Table 1.1. As an example of interpreting the above plots, under the conditions k = k1 and N = N1 then a 1 cm/yr shift in recharge N would change the water-table elevation by slightly more than 45 cm (0.45 m) near the middle of the interfluve.

proportional to ϕ itself. That is, a low ϕ implies a higher sensitivity, and in the specific example plotted the B half of the aquifer has slightly lower ϕ values because river B is 10 m lower in elevation than river A. Thus, while it is not obvious from Fig. 1.3, the maxima of the curves are all slightly shifted towards river B, instead of being at the exact middle of the interfluve. The value $\partial\phi/\partial N$ is also inversely proportional to k; thus aquifers of high permeability are less sensitive than are aquifers of low permeability (Fig. 1.3), as might have been inferred from Fig. 1.1. Hidden within equation 1.3 is the inverse proportionality between $\partial\phi/\partial N$ and N (Fig. 1.3): ϕ, or actually ϕ^2, is directly proportional to N (see equation 1.1), but $\partial\phi/\partial N$ is inversely proportional to ϕ. The intuitive explanation is that as N becomes larger, a unit change in N becomes a smaller proportion of the total N, and hence has less effect in altering ϕ.

DISCUSSION: APPLICATION OF POSITIONAL SENSITIVTY

Meaning and derivation of application

The application of the above equations describing positional sensitivity means to find a $\Delta\Phi$ or $\Delta\phi$ given a ΔN, or vice versa. That is, given a shift in recharge N, how much will the water-table elevation change, and how much will the lake levels associated with that water table change? Conversely, how much change in recharge would be necessary to produce some observed change in water-table elevation or lake level? The change in groundwater potential Φ from state 1 to state 2 (denoted by subscripts) is

$$\Phi_2 - \Phi_1 = \Delta\Phi = (\partial\Phi/\partial N)\,\Delta N = L^2\,S_N\,\Delta N \qquad (1.4)$$

This equation gives an exact value for $\Delta\Phi$ because $\partial\Phi/\partial N$ is independent of both ϕ and N. With the relationship $\Phi = 0.5\,k\,\phi^2$, equation 1.4 may be expressed in terms of the water-table head:

$$\phi_2 - \phi_1 = \Delta\phi = \{(2/k)(L^2\,S_N\,\Delta N + \Phi_1)\}^{0.5} - \phi_1 \qquad (1.5)$$

This equation is also exact, because it is just a rearrangement of equation 1.4. However, equation 1.5 is somewhat cumbersome, and for small changes in N or ϕ then

$$\Delta\phi \approx (\partial\phi/\partial N)\,\Delta N = \{(L^2\,S_N)/(\phi_1\,k)\}\,\Delta N\,C \qquad (1.6)$$

(Again, the constant C merely allows the use of N in units of cm yr^{-1}; if N is in units of m s^{-1}, then C must be omitted.) This equation for $\Delta\phi$ is only approximate, because $\partial\phi/\partial N$ depends on ϕ (and thus also on N). However, the error in using equation 1.6 (instead of equation 1.5) will be less than 10% as long as the resulting estimated change in head ($\Delta\phi_{est}$) satisfies the following inequality:

$$-0.18\,\phi_1 \le \Delta\phi_{est} \le 0.22\,\phi_1 \qquad (1.7)$$

Procedure for application

The above equations strictly apply only to the original simple system modeled by equation 1.1, namely a strip of aquifer lying between two infinitely long parallel rivers.

While such a system does not exist in reality, situations can be found in the field that approximate this simple system. That is, relatively homogenous surficial aquifers lying between two rivers are common occurrences on the landscape, and the trick is to fit the actual geometry, geology, and hydrology of the real landscape to the parameters required by the simple-system model.

As a general example, assume that the landscape includes a small closed-basin water-table lake lying in a relatively homogeneous interfluve, and that one is interested in determining how much the lake level will change if recharge changes by a given amount. First, the geometry of the real landscape must be fit to the simple model. The elevation of the base of the aquifer will have to be determined, *e.g.* the bedrock elevation or, for a sandplain overlying till, the interface of the sand and till. Then, the heads of the rivers (ϕ_A and ϕ_B) and the head at the lake (ϕ_1) may be calculated by subtracting the elevation of the aquifer base from the elevations of these features as indicated on topographic maps. Next, the interfluvial distance is measured. Because no real pair of rivers is truly parallel, this distance is somewhat arbitrary; I suggest measuring the shortest distance from the lake to each river (D_A and D_B), and then letting $2L = D_A + D_B$. Once D_A and L are determined, a value for S_N may be determined from equation 1.2, or a value of S_N may be easily estimated from Fig. 1.2.

Second, the geology and hydrology of the real landscape must be determined; *viz.* values for hydraulic conductivity k and recharge N must be determined. Neither of these parameters is easy to measure, and large errors are to be expected, sometimes up to an order of magnitude (Gillham and Farvolden, 1974; see also Winter, 1981). The best estimates of k and N will come from whatever measurements are available from the actual aquifer under consideration, although rough guesses for k and N could be formulated from fairly scant information about the watershed. For example, N could be estimated from the baseflow of river A or B and from estimating the area enclosed by the groundwater divides surrounding the river. Then k could be solved for from equation 1.1, which would fit a k value such that the modeled water-table head would match that of the lake being investigated. (The next section of the paper will demonstrate the value of using such a fitted k value.)

At this point equation 1.6 may be solved: a ΔN may be chosen, and a $\Delta\phi$ may be calculated, or vice versa. The result calculated in the above manner must be viewed as being very limited and really only semi-quantitative in manner. The result is limited geometrically by the mis-match of the simple-system model with the real landscape: in the model, the water table is heavily constrained by the two rivers extending to infinity in two directions. In real landscapes, rivers are finite, and the water table may move somewhat more freely; hence, except in unusual circumstances, for geometric reasons the simple-system model should *underestimate* a $\Delta\phi$ for a given ΔN. Conversely, the model would *overestimate* a ΔN for a given $\Delta\phi$. As mentioned above, k and N can only be known very approximately, and errors in k and N would propagate through to the calculated $\Delta\phi$ or ΔN value. There is also error in estimating elevations from topographic maps; such elevations are only accurate to within about one-half of the contour interval (Robinson *et al.* 1978). Because of all these limitations, there is probably no reason to use the more exact equation 1.5 instead of equation 1.6.

Example of application: Parkers Prairie sandplain, west-central Minnesota

A sandplain roughly 15 x 15 km surrounds the town of Parkers Prairie in west-central Minnesota (Fig. 1.4). This sandplain was chosen for several important reasons. First, sandplains in general conform well to the assumptions required by two-dimensional groundwater models (see below). Second, this sandplain contains many closed-basin water-table lakes. Investigation of sediments from these lakes has indicated significant lake-level changes in the past in a pattern that is consistent with the above results of positional sensitivity; namely, lake levels (and hence also water-table levels) changed most in those parts of the sandplain farthest away from rivers (Digerfeldt, Björck, and Almendinger, unpublished data; see Acknowledgments). Third, the geologic and hydrologic parameters required for groundwater modeling had already been estimated in an earlier study by the U. S. Geological Survey (McBride, 1975). Fourth, overland runoff and interflow are minimal on a sandplain and hence do not obscure the general agreement of water-table and lake elevations.

The most genuine way of testing the applicability of the simple-system model results to a real aquifer would be to find an interfluvial aquifer for which changes in water-table elevation and changes in recharge had actually been measured, and then to compare these measurements to the $\Delta\phi$ values estimated from equation 1.5 or 1.6. However, obtaining such real measurements is elusive. The next best test is to construct a full groundwater model of an interfluvial aquifer, and to compare the sensitivity of the water table in the full model to the sensitivity indicated by the simple-system equations. Because the full groundwater model requires the same general assumptions as the simple-system model, the comparison really only tests how well the simple-system geometry can be fit to a real landscape.

I constructed a full groundwater model with the analytic-element technique of Strack (1988) (Fig. 1.5). Assumptions include uniform recharge and hydraulic conductivity within each bounded region of substrate type (sand and till), isotropy, and the Dupuit-Forchheimer relation ($\partial\phi/\partial z = 0$, and no resistance to flow in the vertical direction). The aquifer base was placed at the base of the sand and assumed to be horizontal. The boundary of the sandplain was taken from soils maps (Arneman *et al.*, 1969); the locations and elevations of various lakes and rivers were taken from U. S. Geological Survey quadrangle maps. I included major streams and rivers up to a radius of about 75 km away from the center of the sandplain in order to get a broad view of the water table over a large region; increasing detail was mapped toward the center of the model, the portion of interest being only of about 7 km radius. This analytic-element model is of infinite area and thus contains no boundary that circumscribes the region of interest, as opposed to a conventional finite-element or finite-difference model. However, for practical reasons the model is really only accurate in its very central portion, namely, the sandplain itself. The N/k of the model was adjusted until a

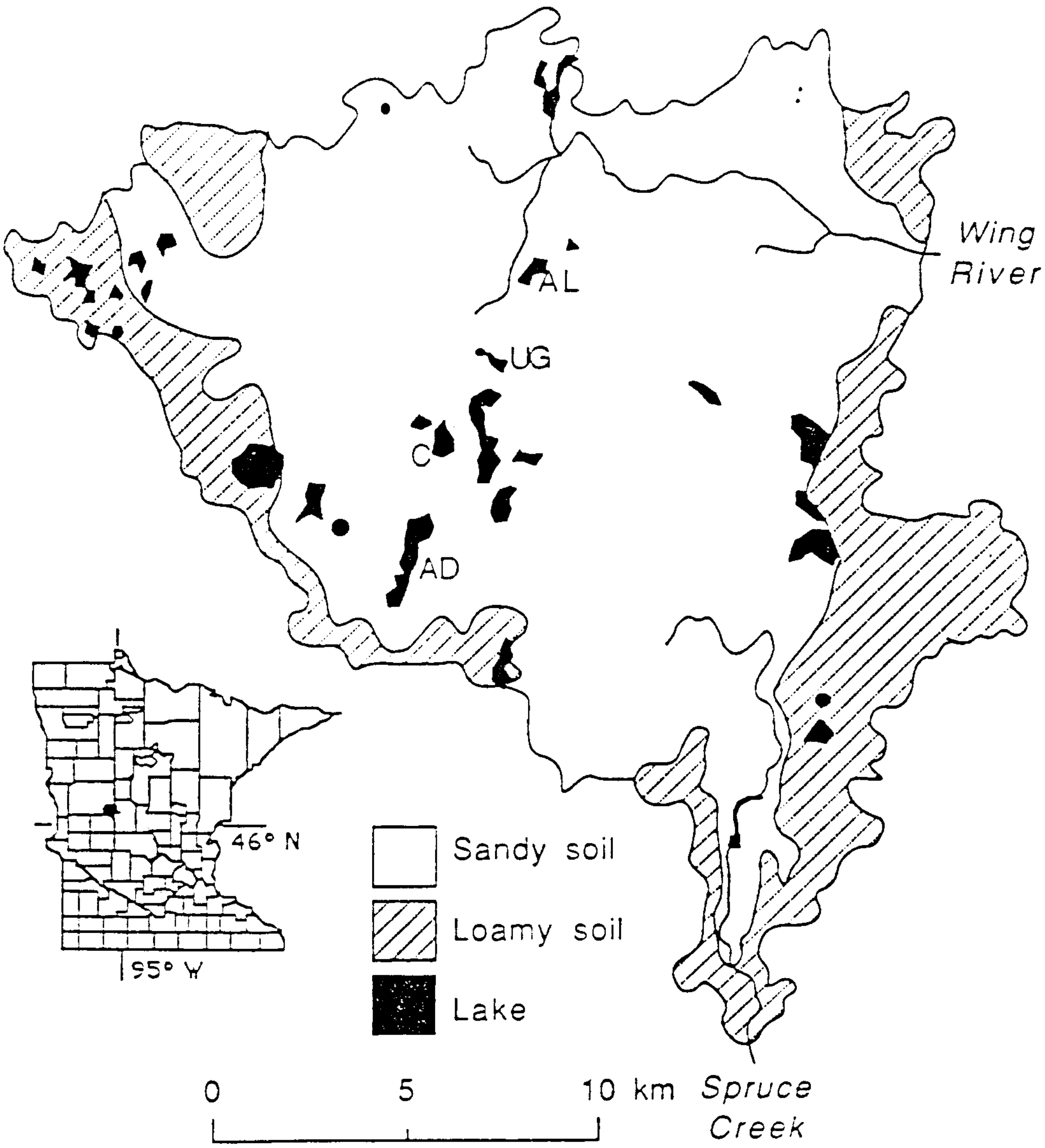

Fig. 1.4. Map of the Parkers Prairie sandplain area: the specific region shown includes the combined watersheds of the Wing River and Spruce Creek. Soil boundaries were taken from Arneman *et al.* (1969); regions of peat within the sandplain are not shown. AL = Almora Lake, UG = Upper Graven Lake, C = Cora Lake, and AD = Lake Adley. The town of Parkers Prairie lies just north of Lake Adley.

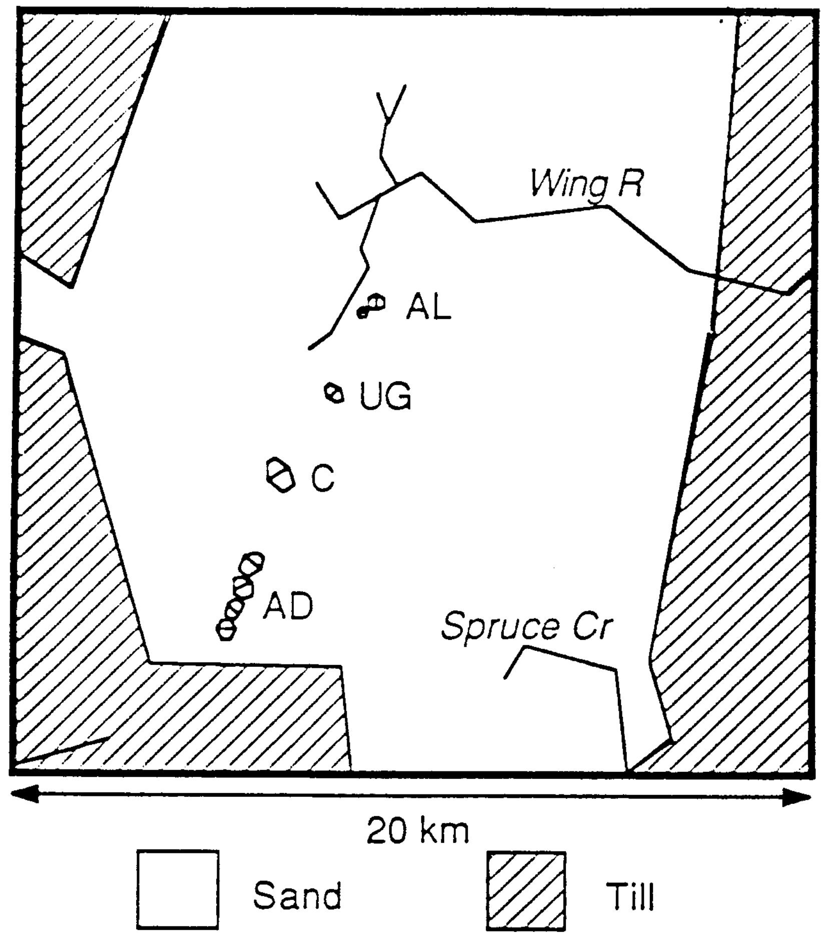

Fig. 1.5. Layout of the central portion of the full groundwater model; model elements extend far beyond the edges of this window (see text). AL = Almora Lake, UG = Upper Graven Lake, C = Cora Lake, and AD = Lake Adley. In this view lakes are shown as combinations of quadrilaterals that are capable of modeling net recharge or discharge over the area of the lake. However, for the model runs in this paper the strength of these elements was set to zero, and hence they had no effect on lake level. Lake levels were taken to be the water-table elevation at the center of each lake basin, or the average elevation for multi-basin lakes.

reasonable fit of the water table was obtained in the sandplain. Then values of N and k were chosen to match as closely as possible those values estimated by McBride (1975); the final full-model value of hydraulic conductivity k_o = 2.25E-3 m s^{-1}, and that of recharge N_o = 5E-9 m s^{-1}, or 15.8 cm yr^{-1}.

Using the procedure outlined in the above section, I fit the simple-system model to the centers of four lakes in the sandplain: Almora, Upper Graven, Cora, and Adley (see Fig. 1.4). For each lake, D_A was the shortest distance to the Wing River, and D_B was the shortest distance to Spruce Creek. I used two different k values: (1) k_o, the same k value used in the full model, and (2) k_{fit}, a unique k value fit to each lake using equation 1 and the estimate of N_o from the full model. In most cases, $k_o > k_{fit}$, a consequence of the geometric insensitivity of the water table of the simple model. That is, the full model is more sensitive than the simple model to N because the simple-model water table is constrained by infinitely long rivers. However, the simple model can compensate for its geometric insensitivity by using a lower k value, which increases the simple-model sensitivity. Thus, the k_{fit} value is not the actual k value of the aquifer but is an effective k value that increases the simple model sensitivity enough to allow a fit of the simple-model water table and elevations of the two rivers and lake in question.

The full model was run at various proportions of N_o (0.2, 0.4, 0.6, 0.8, 1.0, 1.5, and 2.0 times N_o), and for each run a lake-level change was calculated for each of the four chosen lakes; by definition, lake-level change was zero for the 1.0 N_o run. Then, for each of these values of recharge, I used the simple model, as fit to each lake, to estimate a lake-level (water-table elevation) change. In this case, I used the more-exact equation 1.5 rather than the approximate equation 1.6. I ran the simple model using both k_o (the k of the full model) and k_{fit}, as explained above.

The plots of the full-model and simple-model estimates of lake-level change as a function of recharge (Fig. 1.6) made evident several points. First, and most importantly, in all comparable model runs Lake Adley (Fig. 1.5D) was the most sensitive lake, specifically because this lake is located the farthest from the rivers draining the sandplain. Cora Lake is next farthest, followed by Upper Graven and Almora (see Fig. 1.4); in all cases lake-level sensitivity followed the same order and was thus directly proportional to the distance from the lake to the nearest river. Second, as expected, the simple model generally underestimated lake-level sensitivity, particularly when the same hydraulic conductivity k_o was used in the simple model as in the full model. However, when k_{fit} was used instead, the simple model regained enough sensitivity to produce lake-level changes very similar to those of the full model. Third, at very low values of recharge the rivers in the sandplain may begin to dry up. As the rivers shorten, they get farther away from the lakes, and hence the lakes become even more sensitive. The full model could account for the shortening of the rivers, but the simple model could not. Hence the simple model began to greatly underestimate lake-level changes at very low recharge values.

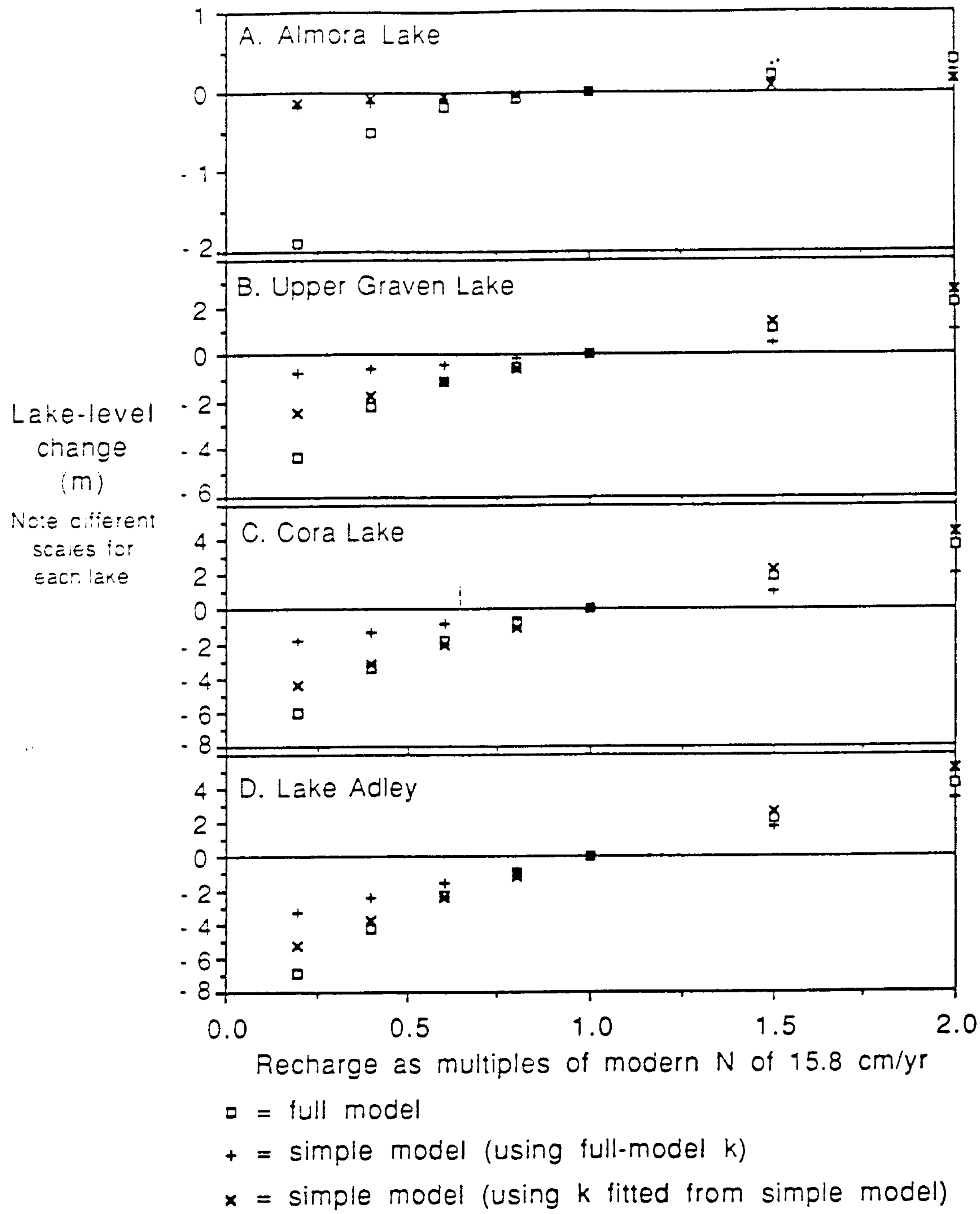

Fig. 1.6. Comparisons of the lake-level changes predicted by the full groundwater model with those predicted by the simple model (equation 1.5), in response to a given change in recharge N. Plots are in the order of the distances of the lakes from the Wing River, as depicted in Fig. 1.4.

SUMMARY AND CONCLUSIONS

Given a change in recharge to a surficial aquifer under steady-state conditions, closed-basin lakes far from a river will change their levels more than will lakes close to a river, and minimum values for such lake-level changes may be estimated from a simple analytic model.

These results lead to some important conclusions for the study of both modern and paleo-lake levels. In modern systems it seems clear that lakes lying farthest from rivers are the most vulnerable to a lake-level change caused by a change in recharge. For example, School Section Lake (mentioned in the introduction) sits near the crest of an interfluvial groundwater divide (Mohring, 1986), a very sensitive position. Further, application of the positional sensitivity of a lake (as fit to the simple model) may help provide a quick, althouth very rough, estimate of the minimum expected lake-level (and water-table) change resulting from a given change in recharge.

For paleoecological work, the understanding of the positional sensitivity of lakes should help first of all in selection of sites for study. If the goal is to find a stable lake with an uninterrupted sedimentary record, then a lake with an outlet, or at least near a river, would be a likely candidate. However, if the goal is to investigate past lake levels as an indicator of climatic change, then lakes far from rivers would be good choices. The concept of positional sensitivity is also very important in the interpretation of paleo-lake levels. Because of the different positions of lakes in a watershed, a uniform shift in climate can cause widely different magnitudes of lake-level changes, which could be misinterpreted if only the surficial hydrology of the lakes is considered. Further, the simple understanding that groundwater recharge can affect water tables and lake levels should help provide some realistic complexities in the interpretation of past lake levels. A shift in the paleo-recharge rate may be the result of change in some mean-annual climatic parameter, but recharge can also change abruptly in response to a change in vegetation.

CHAPTER 2

GROUNDWATER CONTROL OF CLOSED-BASIN LAKE LEVELS: 2. HOW RIVER PROXIMITY CHANGES THE EFFECT OF EVAPORATION AND DIRECT PRECIPITATION UNDER STEADY-STATE CONDITIONS

ABSTRACT

"Lake pumping" is defined as the net removal of water from a lake by hydrologic processes acting at the lake surface, namely evaporation minus direct precipitation and minus any input from overland runoff that reaches the lake. Investigation of a simple analytic groundwater model of a circular lake next to an infinitely long river shows that the sensitivity of the lake level to a change in lake pumping is proportional to the distance from the lake to the river. However, the river has little power in constraining lake level unless the lake is close to the river, *i.e.* within a distance of five or ten lake radii. Beyond that distance, large lakes are more sensitive than small lakes. The analysis also indicates that lakes lying in highly permeable substrates are not very sensitive to changes in lake pumping. These results may help interpret the causes of paleo-lake-level fluctuations revealed by geologic investigations.

INTRODUCTION

Levels of closed-basin lakes have been known for over 250 years as indicators of climate (Smith and Street-Perrott, 1983). Such lakes may also be sensitive to hydrologic shifts effected by non-climatic factors such as changes in the vegetation or changes in land use by man. On the modern landscape lake-level change can cause property damage and devaluation, and understanding the factors to which these lake levels are sensitive would be valuable. Further, the inference of past climate from knowledge of paleo-lake levels can be an important tool in the testing or calibration of climate models (Kutzbach, 1985). Previous studies relating past climate to paleo-lake levels have been limited to sites where groundwater flow was assumed to be simple enough to preclude the need for modeling it explicitly (*e.g.*, Kutzbach, 1980; Hastenrath and Kutzbach, 1983; Street-Perrott and Harrison, 1985). Knowledge of how groundwater processes may affect the sensitivity of closed-basin lake levels would increase the number of sites amenable to paleoclimatic investigation, as well as provide new insights on the details of past climate.

Closed-basin lakes hydraulically connected with the water table may change their levels by two fundamental mechanisms. First, lake levels generally track fluctuations in the elevation of the *regional* water table. Such fluctuations may result from shifts in groundwater recharge as described in Chapter 1. River downcutting, or damming of streamflow by peat growth, for example, may change the elevation of the stream draining the aquifer; such a change in stream elevation could likewise have a regional or sub-regional effect on the water-table elevation.

The second mechanism of changing closed-basin lake levels is the *local* process of directly removing water from the lake itself via surficial means, or adding water to it.

That is, evaporation E directly removes water from the lake ("surficial output"), and precipitation P on the lake surface, overland runoff, and perhaps interflow directly add water to the lake ("surficial inputs"; all units must be consistent, *e.g.* as cm yr^{-1} over the lake surface, not catchment). Under steady-state conditions, the lake responds only to the net amount of water removed (or added), and the lake cannot recognize the absolute amounts of the individual components of surficial outputs and inputs. I define "lake pumping rate," symbolized by γ, as the net areal discharge from the lake surface, namely surficial outputs minus surficial inputs. In a sandplain, overland flow and interflow are minimized, and $\gamma \approx E - P$.

Lakes are three-dimensional depressions in the landscape that generally intersect the water table, and the groundwater flow patterns around and below a lake may be complex (Winter, 1976, 1978; Pfannkuch and Winter, 1984; Winter and Pfannkuch, 1984). However, if the aquifer is thin compared with its breadth nearly all of the groundwater flow will be two-dimensional in the horizontal plane, and closed-basin lakes will appear as large-diameter wells, with groundwater entering or exiting through the lake margin much as through a well screen. Indeed, McBride and Pfannkuch (1975) used both theoretical and field evidence to demonstrate that, for geometric reasons, most of the groundwater seepage into a lake tends to be concentrated right along the lake margin. Thus many lakes, at least to *some* degree, may be considered as large-diameter wells that pump water out of the aquifer via evaporation and pump water into the aquifer via precipitation falling on the lake surface. Under steady-state conditions the net volumetric rate of water pumped out of a lake would be equal to the lake pumping rate γ multiplied by the surface area of the lake. Note that γ may be negative (if, *e.g.*, $P > E$), in which case the lake will act as a large-diameter injection well.

My purpose is to investigate how the aquifer surrounding a lake can modify the efficacy of surficial hydrology (collapsed into the single term "lake pumping" γ) in changing lake levels. To this end I examine a very simple analytic groundwater model of a circular lake next to an infinitely long river. Being forced to maintain constant head, the river tends to constrain any lake-level change caused by lake pumping, and consequently the farther lakes are from the river the more sensitive they are to lake pumping. However, the river's constraining power diminishes rapidly with distance so that only lakes quite near the river are insensitive to lake pumping. Outside the range of the river's influence, large lakes are more sensitive to lake pumping than are small lakes, as one might intuitively expect. I then apply the above simple model results to lakes in a sandplain in west-central Minnesota, and find that when lakes and rivers are placed in a geometrically more realistic pattern the river has even less control on lake levels than in the simple model.

METHOD: SIMPLE-SYSTEM EQUATION FOR GROUNDWATER POTENTIAL

To simplify the examination of lake-level sensitivity with respect to lake pumping, I chose a system of very simple geometry. For a circular lake next to an infinitely long river, the groundwater potential Φ along the lake margin is given by

$$\Phi = \Phi_o - (\gamma r^2/2) \ln\{[D + (D^2 - r^2)^{0.5}]/r\} \qquad (2.1)$$

For an unconfined aquifer, groundwater head $\phi = (2\,\Phi/k)^{0.5}$. (See Fig. 2.1 and Table 2.1 for an explanation of parameter symbols and values; see the appendix for a derivation of equation 2.1.) Equation 2.1 is a modification of a simple system in which a well and image well are separated by a linear equipotential; there is no uniform flow or flow from infinity, and hence Φ approaches Φ_o at infinity. These wells are surrounded by circular eccentrically placed equipotentials, and one of these equipotentials may be chosen as a lake boundary. This equation models two-dimensional groundwater flow in the horizontal plane and assumes homogeneity and isotropy of the aquifer, as well as the Dupuit-Forchheimer relation ($\partial\phi/\partial z = 0$, and no resistance to flow in the vertical direction). The water-table head at the lake boundary is proportional to the quantity $-\gamma/k$, meaning that a smaller k or a greater γ lowers the lake level (and the surrounding water table), as shown schematically in Fig. 2.1B.

RESULTS: LAKE-LEVEL SENSITIVITY

Lake levels change as a result of lake pumping because of the volumetric rate Q ($m^3\,s^{-1}$) at which water is being removed or added. For a circular lake of radius r (m) undergoing a lake pumping rate of γ ($m\,s^{-1}$), $Q = \gamma\,\pi\,r^2$. Lake level is thus responsive to both γ and r, and increasing either γ or r increases Q and hence tends to lower lake level. The sensitivity of the lake level to these two parameters is just the partial derivative of the groundwater head at the lake boundary (given by equation 2.1) with respect to either γ or r. Note that in the following analyses, lake-level sensitivity will generally be measured in the *negative* direction, *i.e.*, the "greater" sensitivities will be the more negative sensitivities when calculated as partial derivatives. Because γ was defined as the lake's net areal discharge, as opposed to recharge, an increase (a positive change) in γ or r will tend to cause a decrease (a negative change) in lake level.

Sensitivity with respect to lake radius

S_r: the dimensionless sensitivity with respect to lake radius

Equation 2.1 may be differentiated with respect to r, holding γ constant. To make the result independent of the dimensions of any specific aquifer and lake, all linear distances were normalized by D, and all rates were normalized by γ. The resulting derivative is

$$S_r = (\partial\Phi/\partial r)/(\gamma\,D) = (k\,\phi\,\partial\phi/\partial r)/(\gamma\,D) =$$
$$= (-r_D/2)[2\ln\{(1 + (1 - r_D^2)^{0.5})/r_D\} - 1/(1 - r_D^2)^{0.5}] \qquad (2.2)$$

Table 2.1. Definitions of parameter symbols and values of constants

Symbol	Definition or value (unit of measurement)
ϕ	= water-table head (m)
ϕ_o	= head at river = 20 m
k	= hydraulic conductivity (m s^{-1})
k_1	= 1E-3 m s^{-1}, coarse sand and gravel
k_2	= 0.5 k_1 = 5E-4 m s^{-1}
Φ	= 0.5 k ϕ^2 = groundwater potential (m^3 s^{-1})
Φ_o	= 0.5 k ϕ_o^2 = groundwater potential at river (m^3 s^{-1})
γ	= areal discharge from the lake, or "lake pumping" (m s^{-1}, or cm yr
γ_1	= 2E-8 m s^{-1}
γ_2	= 2 γ_1 = 4E-8 m s^{-1}
Q	= $\gamma \pi r^2$ = volumetric rate of lake pumping (m^3 s^{-1})
D	= distance from river to center of lake (m)
r	= lake radius (m)
D_r	= D/r = distance from river relative to r (unitless)
r_D	= r/D = lake radius relative to D (unitless)
S_r	= $(\partial\Phi/\partial r)/(\gamma D)$ = $(k\,\phi\,\partial\phi/\partial r)/(\gamma D)$ = dimensionless sensitivity of the groundwater potential at the lake boundary with respect to a change in lake radius, holding lake pumping constant (unitless)
$\partial\phi/\partial r$	= $\{(\gamma D)/(k\,\phi)\}\,S_r$ = actual sensitivity of lake level with respect to a change in lake radius, holding lake pumping constant (unitless)
S_γ	= $(\partial\Phi/\partial\gamma)/r^2$ = $(k\,\phi\,\partial\phi/\partial\gamma)/r^2$ = dimensionless sensitivity of the groundwater potential at the lake boundary with respect to a change in lake pumping, holding lake radius constant (unitless)
$\partial\phi/\partial\gamma$	= $\{(C\,r^2)/(k\,\phi)\}\,S_\gamma$ = actual sensitivity of lake level with respect to a change in lake pumping, holding lake radius constant (m/(cm yr^{-1}), or m/(m s^{-1}) if C is omitted)
C	= 3.171E-10 (m s^{-1})/(cm yr^{-1}), merely a conversion factor

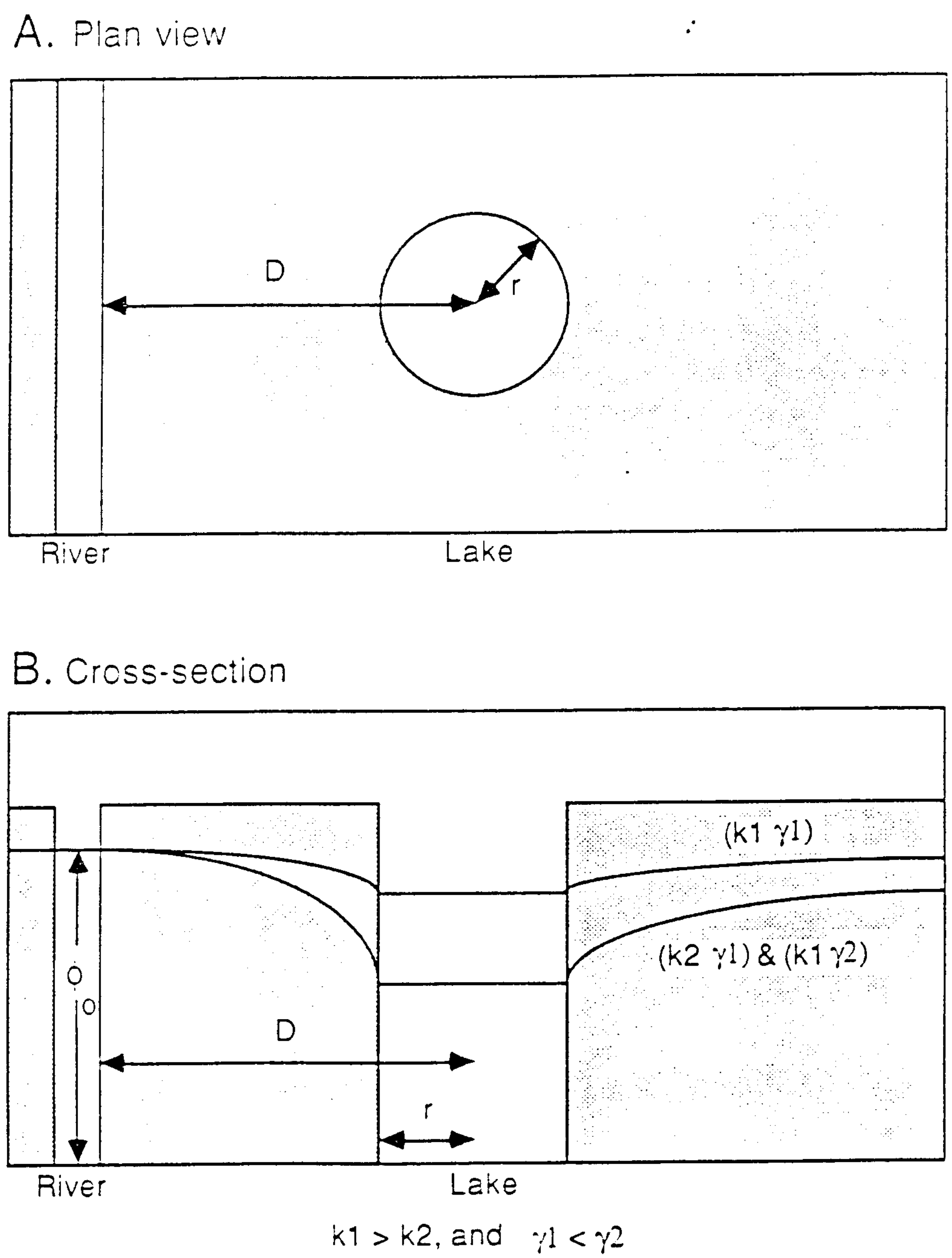

Fig. 2.1. Plan view (A) and cross-section (B) of the simple system of a circular lake beside an infinitely long river. This figure is only schematic: equation 2.1 may be used to calculate the groundwater potential Φ or head ϕ at the boundary of the lake. Definitions of symbols are given in Table 2.1.

in which $r_D = r/D$ (see Table 2.1). S_r is the dimensionless sensitivity of the groundwater potential with respect to lake radius and is a function of only the relative size (radius) of the lake.

Equation 2.2 (as depicted in Fig. 2.2) demonstrates the two contrasting controls on the lake level: (1) when positive (*i.e.*, when $E > P$), γ tends to *lower* lake level, and (2) the river tends to *maintain* lake level at the river elevation. That is, in general, large lakes have a lower level than small lakes given the same γ, unless the lakes are (relatively) close to the river, at which point the constraining effect of the river overrides any lowering tendency caused by γ. Specifically, as long as $r_D < 0.7617$, S_r is less than zero, meaning that increasing the lake radius will lower the lake level. The maximum sensitivity occurs at $r_D = 0.3584$; *i.e.*, when the lake is that size, a unit increase in lake radius will lower the lake more than when the lake is any other size. When $r_D > 0.7617$, S_r is positive and the lake level is controlled more by the river elevation than by γ, until ultimately $r_D = 1$ and the lake is tangent to the river, whereupon the lake level is completely controlled, and hence infinitely sensitive, to whatever elevation the river happens to be. The point at which $r_D = 0.7617$ is the critical point of zero sensitivity, at which the lowering tendency of γ is exactly balanced by the constraining tendency of the river. The sensitivity is also zero for infinitesimally small lakes, because a tiny increase in the radius of a tiny lake increases the lake area by an infinitesimal amount, and hence the effect of γ in lowering the lake level cannot be expressed.

$\partial\phi/\partial r$: the actual sensitivity with respect to lake radius

Equation 2.2 may be rearranged as follows:

$$\partial\phi/\partial r = \{(\gamma D)/(k\,\phi)\}\, S_r \qquad (2.3)$$

The quantity $\partial\phi/\partial r$ represents the actual sensitivity of the lake level (= piezometric head at the lake boundary) with respect to lake radius under the specific conditions of γ, D, and k; ϕ may be calculated from equation 2.1.

To evaluate this expression, values must be chosen for γ, D, and k. The plot of equation 2.3 as a function of lake radius (Fig. 2.3A) is morphometrically identical to Fig. 2.2: the two figures have the same critical points along the abscissa. Further, Fig. 2.3A demonstrates the increased sensitivity of lakes under conditions of either lower k or higher γ, as might have been inferred from Fig. 2.1. As an example of interpreting Fig. 2.3A, consider an aquifer with a hydraulic conductivity of k_1, a rate of lake pumping of γ_1, and two lakes (A and B, not close to each other) whose centers are 1 km away from the river. If lake A has a radius of 350 m, and lake B a radius of 351 m, then lake B would be about 0.4 mm lower because of the increased effect of γ on lake B's slightly larger area. This tiny lake-level difference is to some degree an artifact of the very high hydraulic conductivity value (1E-3 m s^{-1}, coarse sand and gravel) I used in this specific example; more common k values would be orders of magnitude less, and the corresponding lake-level differences would be commensurately greater. This

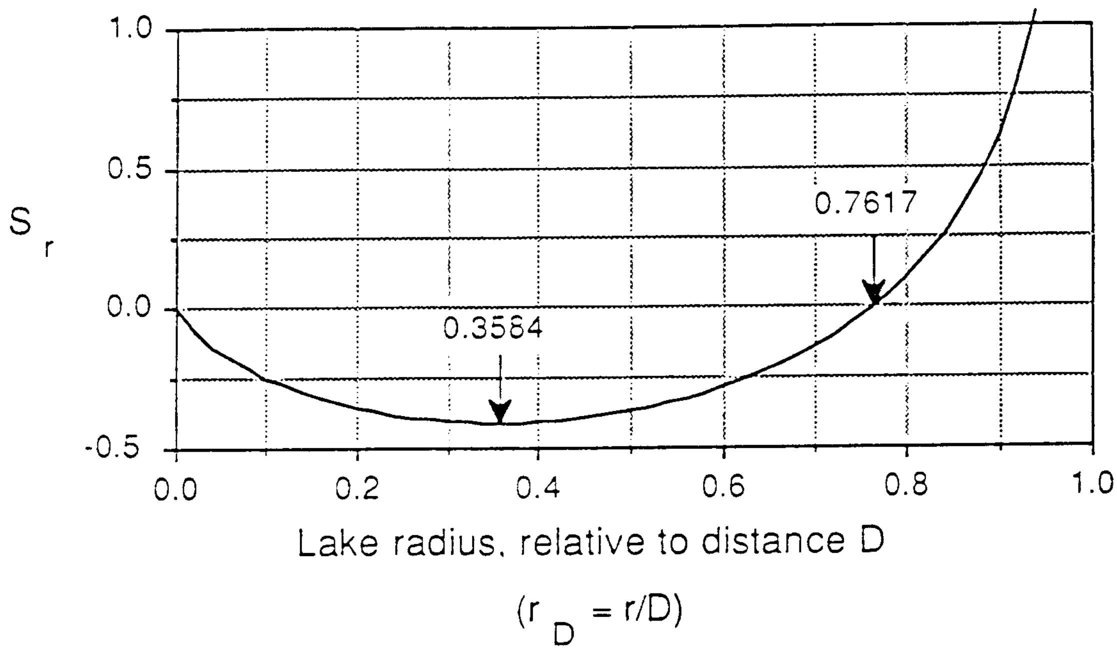

Fig. 2.2. Plot of equation 2.2: S_r, the dimensionless sensitivity of the groundwater potential at the lake boundary with respect to a change in lake radius, holding lake pumping γ constant. See text for explanation of critical points.

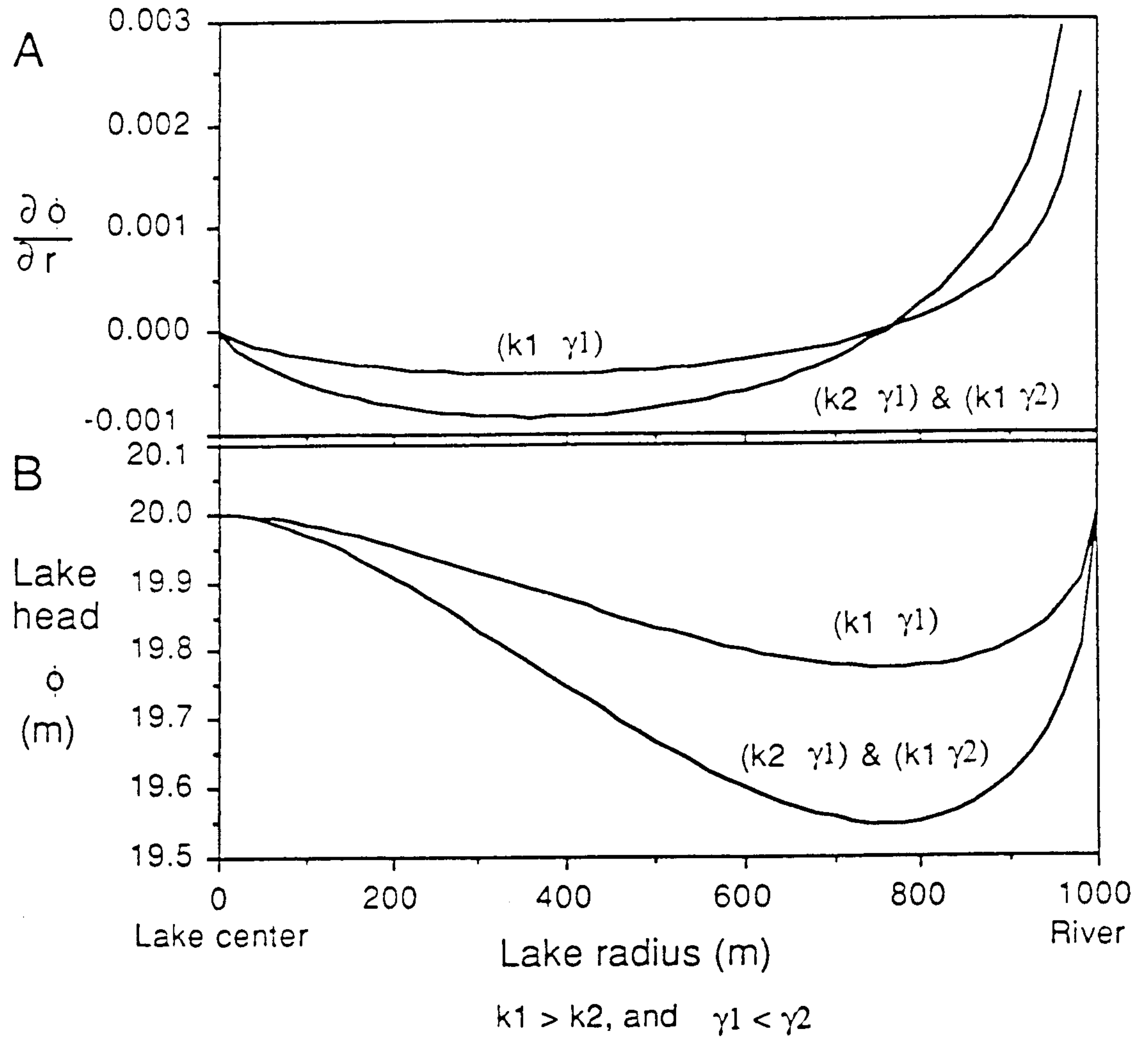

Fig. 2.3. (A) Plots of equation 2.3: examples of the actual sensitivity of the lake level with respect to a change in lake radius, holding lake pumping γ constant, as a function of lake radius. k1 > k2, and $\gamma 1 < \gamma 2$. The distance D between the lake center and the river is held constant at 1000 m in these examples; values for other parameters are given in Table 2.1. (B) Plots of equation 2.1 solved for ϕ: lake level as a function of lake radius, with the values of parameters the same as above.

example does emphasize, however, that lakes lying in highly permeable substrates are not particularly sensitive to either γ or r.

The plot of lake level itself (ϕ) as a function of lake radius (Fig. 2.3B) emphasizes the meaning of the critical points noted in Figs. 2.2 and 2.3A. The steepest negative slope is for a lake of radius 358.4 m ($\Rightarrow r_D = 0.3584$), and the slope is zero for a lake of radius of 761.7 m ($\Rightarrow r_D = 0.7617$), for which size a lake would have the lowest possible level. The plot further emphasizes the insensitivity of lakes lying in highly permeable settings: under conditions of k_1 and γ_1, lake pumping γ could lower lake level at most by only roughly 20 cm.

Sensitivity with respect to γ

S_γ: the dimensionless sensitivity with respect to γ

Equation 2.1 may be differentiated with respect to γ (holding r constant). To make the result independent of the dimensions of the aquifer and lake, all distances were normalized by (= divided by) the lake radius, giving

$$S_\gamma = (\partial\Phi/\partial\gamma)/r^2 = (k\ \phi\ \partial\phi/\partial\gamma)/r^2 = -0.5 \ln\{D_r + (D_r^2 - 1)^{0.5}\} \qquad (2.4)$$

in which $D_r = D/r$, and $D_r \geq 1$ (see Table 2.1 for explanation of parameters). S_γ is the dimensionless sensitivity of the groundwater potential with respect to lake pumping γ, and is a function of only the distance, relative to r, between the center of the lake and the river.

The plot of equation 2.4 (Fig. 2.4) monotonically decreases, starting with a zero sensitivity at the point when $D_r = 1$ and the lake is tangent to the river and completely controlled by it. The sensitivity continues to "increase" (get more negative, actually) without bound as lakes get farther away from the river. However, the steepest increase in sensitivity occurs within a distance of five or ten lake radii, and beyond that distance the sensitivity increases only slowly with increasing distance.

$\partial\phi/\partial\gamma$: the actual sensitivity with respect to γ

In order to apply S_γ, information regarding a specific aquifer, lake, and river must be supplied. Rearranging equation 2.4 gives

$$\partial\phi/\partial\gamma = \{(C\ r^2)/(k\ \phi)\}\ S_\gamma \qquad (2.5)$$

in which C is merely a constant that allows lake pumping γ to be measured in units of cm yr^{-1} (reasonable units in terms of lake evaporation and precipitation) instead of m s^{-1}. The quantity $\partial\phi/\partial\gamma$ is the actual sensitivity of the lake level (= the water table at the lake boundary) with respect to lake pumping γ and is dependent on the specific values of r and k chosen; ϕ may be found via equation 2.1, which further requires that γ be specified.

The plots of $\partial\phi/\partial\gamma$ as a function of the distance between the lake center and the river (Fig. 2.5A) are essentially specific cases of Fig. 2.4. The lake radius was fixed at 250 m, and different values of k and γ were inserted into equation 2.5. The lake-level

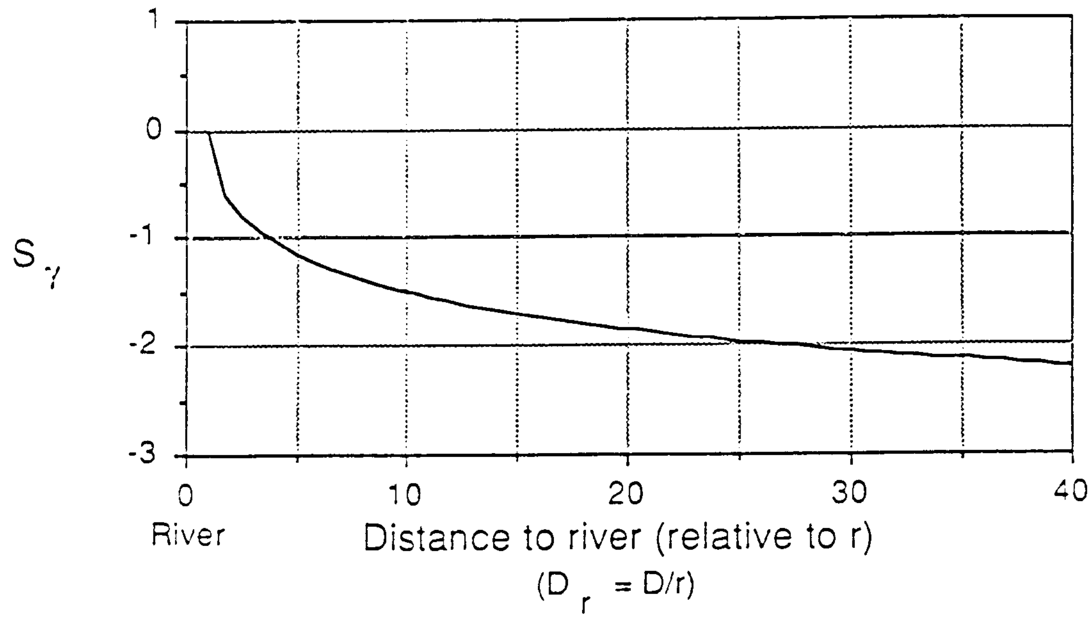

Fig. 2.4. Plot of equation 2.4: S_γ, the dimensionless sensitivity of the groundwater potential at the lake boundary with respect to a change in lake pumping γ, holding lake radius r constant; sensitivity is measured in the negative direction. The plot begins at $D_r = 1$, at which point the lake is tangent to the river.

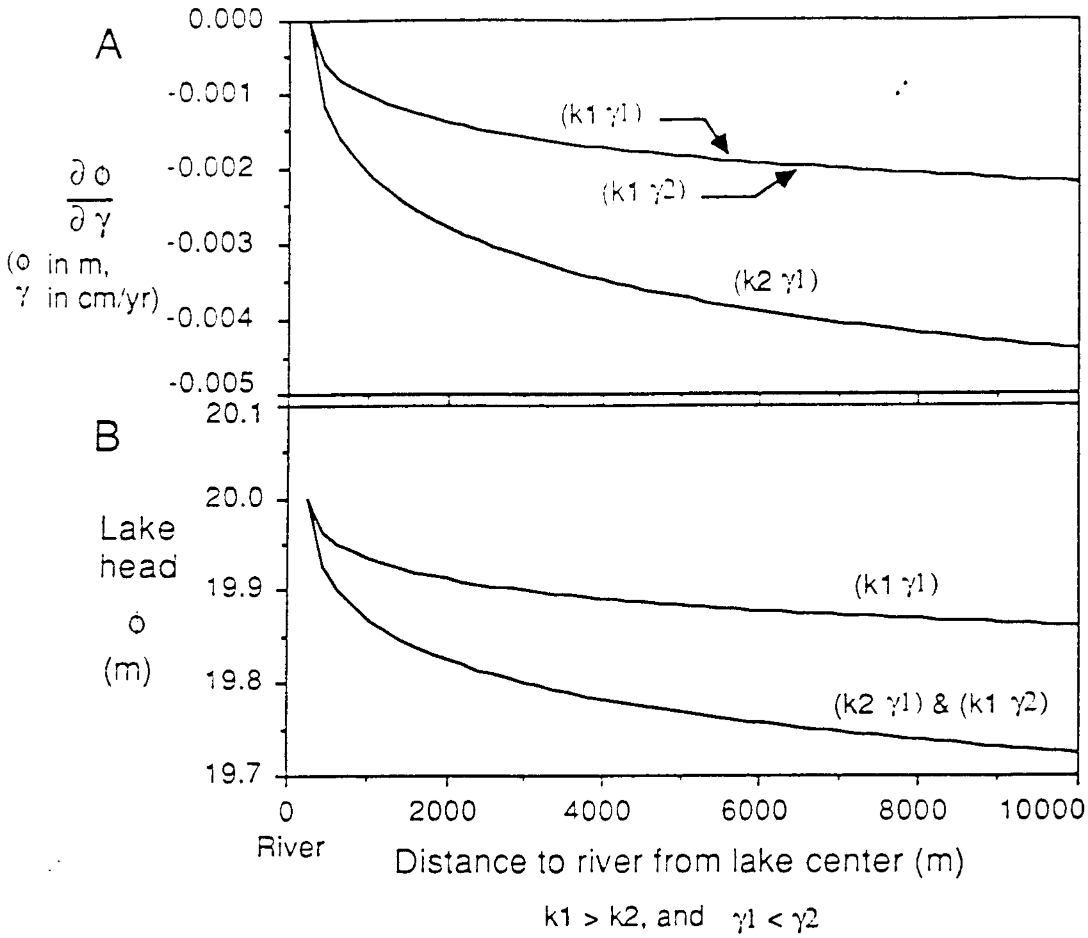

Fig. 2.5. (A) Plots of equation 2.5: examples of the actual sensitivity of the lake level with respect to a change in lake pumping γ, holding lake radius r constant at 250 m; sensitivity is measured in the negative direction. k1 > k2, and $\gamma 1 < \gamma 2$. Values for these and other parameters are given in Table 2.1. Sensitivity is proportional to γ, and hence the sensitivity under conditions k1 $\gamma 2$ is actually slightly greater (= more negative) than that under conditions k1 $\gamma 1$. However, the high value of k1 in this example overwhelms the effect of γ, and the two plots appear to be the same. (B) Plots of equation 2.1: lake level as a function of distance from the river, for a lake of the same size and under the same conditions as above.

sensitivity is proportional to γ and inversely proportional to k; equation 2.5 also implies that $\partial\phi/\partial\gamma$ is also proportional to r^2 (not shown in Fig. 2.5A). As an example of interpreting the plots, consider a lake that sits in an aquifer with a hydraulic conductivity of k_2, is located 4 km from the river, and experiences a lake pumping rate of γ_1. Fig. 2.5A then implies that an increase in γ of 1 cm yr^{-1} would lower the lake level by 3.5 mm. Again, the small change in lake level is a result of the high permeability used in this specific example, and lower permeabilities would cause greater lake-level changes.

Equation 2.1 may be solved for ϕ and plotted as a function of distance between the lake center and the river (Fig. 2.5B) to demonstrate how lake level itself would change under the same conditions of r, k, and γ used in Fig. 2.5A. Fig. 2.5B is not a plot of the water table but rather of the level of a 250-m radius lake located farther and farther away from the constraining river. The morphometry of Fig. 2.5B is qualitatively identical to that of Figs. 2.4 and 2.5A, demonstrating that lakes farther from the river, lakes lying in substrates of lower permeability, and lakes experiencing higher rates of lake pumping are more sensitive to the lake-level change caused by lake pumping. Shown again is the insensitivity of lakes lying in the highly permeable substrates chosen for this specific example: for a lake lying in coarse sand and gravel (k_1 = 1E-3 m s^{-1}) and located 4 km from the river, doubling lake pumping γ (from γ_1 to γ_2) lowers lake level by only 10 cm.

DISCUSSION: THE APPLICABILITY OF LAKE-LEVEL SENSITIVITY TO LAKE PUMPING

The above results strictly apply only to the simple and unrealistic case of a circular lake lying next to an infinitely long river. An important facet to explore is the applicability of the simple-system results to a more realistic system. In other words, what are the biases of the simple system, and how may the results of the simple system be used to gain useful information about real systems?

Any groundwater model is limited by its assumptions. The simple system (equation 2.1) assumes steady state, homogeneity, isotropy, and the Dupuit-Forchheimer relation, and the model will be inaccurate to the degree that these assumptions are violated (such violations are explored in the aforementioned Winter and Pfannkuch papers). While all real aquifers violate all of these assumptions, to some degree the severity of the violations is somewhat scale-dependent, and local complexities do not necessarily invalidate general conclusions gleaned from two-dimensional horizontal models of regional flow systems. In some respects, a regional two-dimensional model is not so much wrong as it is incomplete: the general regional trends are present, but local complexities must be superimposed if detailed information about a specific site is desired. Assume, then, that the simple system modeled by equation 2.1 has some general validity on a regional scale; even then the simple system is limited by its geometry. In the simple system the water table is perfectly constrained along an infinitely long river, and no real system would have a water table so heavily constrained. That is, the simple system is geometrically biased to *underestimate* lake-level sensitivity, and it is expected that a real system would have greater lake-level fluctuations than those predicted by the results of the analysis of the simple system.

Additionally, the simple system can handle only one lake at a time, whereas a real system, or even a regional groundwater model, would include many lakes and peatlands. If, for example, lake pumping (E - P) were highly positive, the total evaporation from many lakes and peatlands across the region could lower the general water table. Thus a lake level would lower not only because of local evaporation from that lake itself, but also because of the total regional evaporation from other lakes and peatlands. The simple system necessarily ignores such a regional effect, and once again the simple system is biased to underestimate lake-level sensitivity when compared to a more realistic system.

The ideal check of applying the simple-system results would be to have actual field measurements of lake-level response to different values of lake pumping, and then to compare these measurements to lake-level changes predicted by fitting the simple model to these lakes. Such actual measurements are not available, and the next best test would be to compare the lake-level changes predicted by the simple system to those predicted by a full, regional groundwater model of a real aquifer. For this comparison I chose the Parkers Prairie sandplain in west-central Minnesota (Fig. 2.6), and used essentially the same groundwater model as described in Chapter 1. In this version of the model (Fig. 2.7), the lakes were modeled as simple rectangular elements over which a lake pumping rate (E - P) could be imposed. The full model requires the same general assumptions as the simple model (see Chapter 1), and the comparison of the predicted lake-level changes primarily demonstrates the limitations of the simple system's geometry versus the more realistic geometry of the full model.

Three lakes in the sandplain were chosen for comparison of the simple and full groundwater models: Almora Lake, Upper Graven Lake, and Lake Adley (Figs. 2.6 and 2.7). The simple model was fit to each of these lakes by calculating an effective radius from the known area of each lake, and by using the distance between the lake and the nearest point on the Wing River as the distance D (see equation 2.1). As expected, the simple model greatly underestimated the lake-level change caused by lake pumping as calculated in the full groundwater model (Fig. 2.8). Not only is the Wing River not infinite in length, but the lakes lie off the very tip of the river; hence the lakes are much less constrained geometrically in the full model than in the simple model. Thus even in the simple model, the river's constraint on lake levels is limited to the near-river region; in a system of more realistic geometry the river's power of constraint is even less.

SUMMARY AND CONCLUSIONS

(1) The larger a closed-basin lake is, the more sensitive it is to a constant rate of lake pumping γ, except when the lake is extremely close to a river, *e.g.*, when the distance between the edge of the lake and the river is about half the lake's radius. (2) The sensitivity of closed-basin lakes to a changing rate of lake pumping γ is proportional to the distance to a constraining river; however, most of the river's constraint on lake

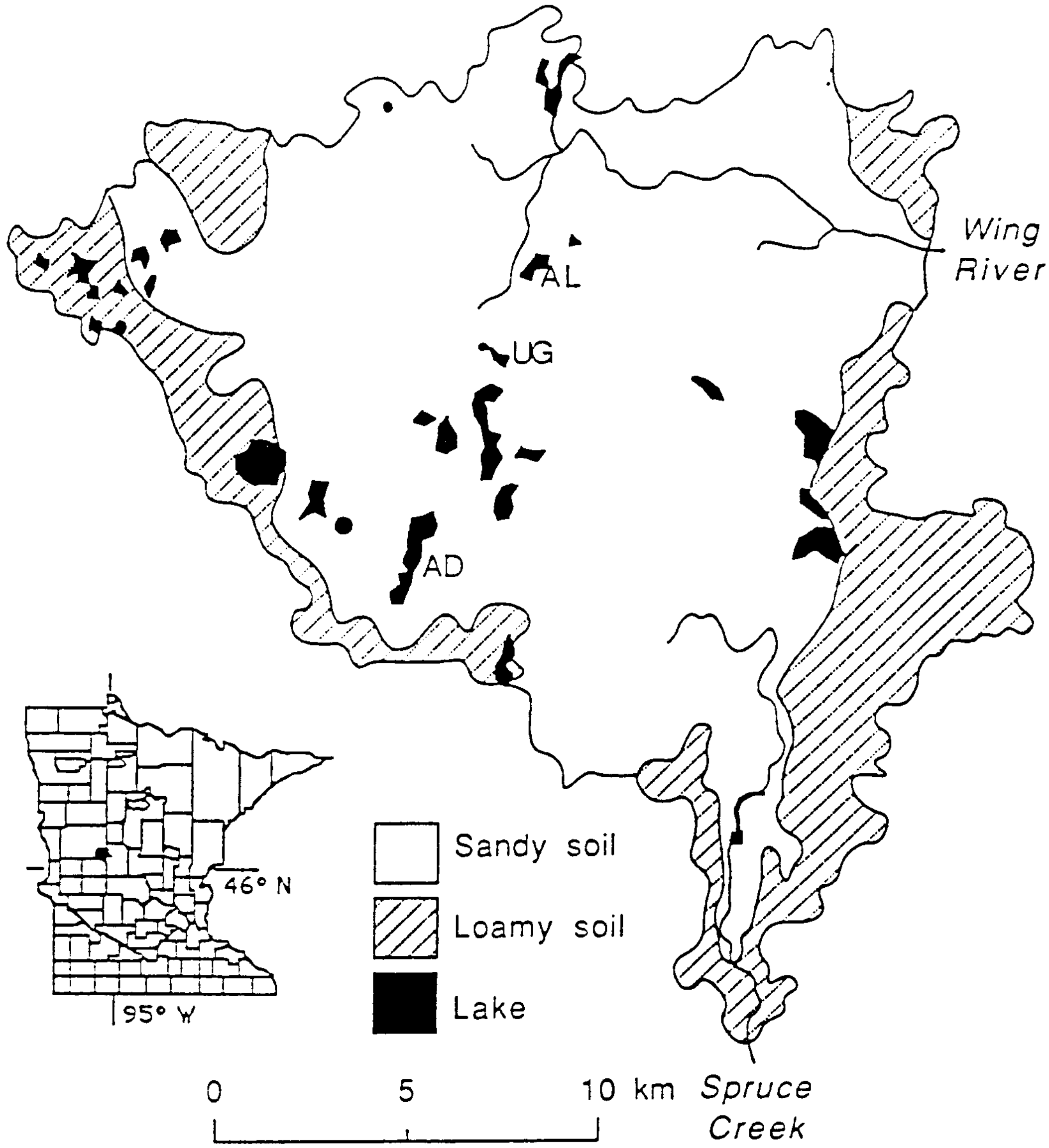

Fig. 2.6. Map of the Parkers Prairie sandplain area; the specific region shown includes the combined watersheds of the Wing River and Spruce Creek. Soil boundaries were taken from Arneman *et al.* (1969); regions of peat within the sandplain are not shown. AL = Almora Lake, UG = Upper Graven Lake, and AD = Lake Adley. The town of Parkers Prairie lies just north of Lake Adley.

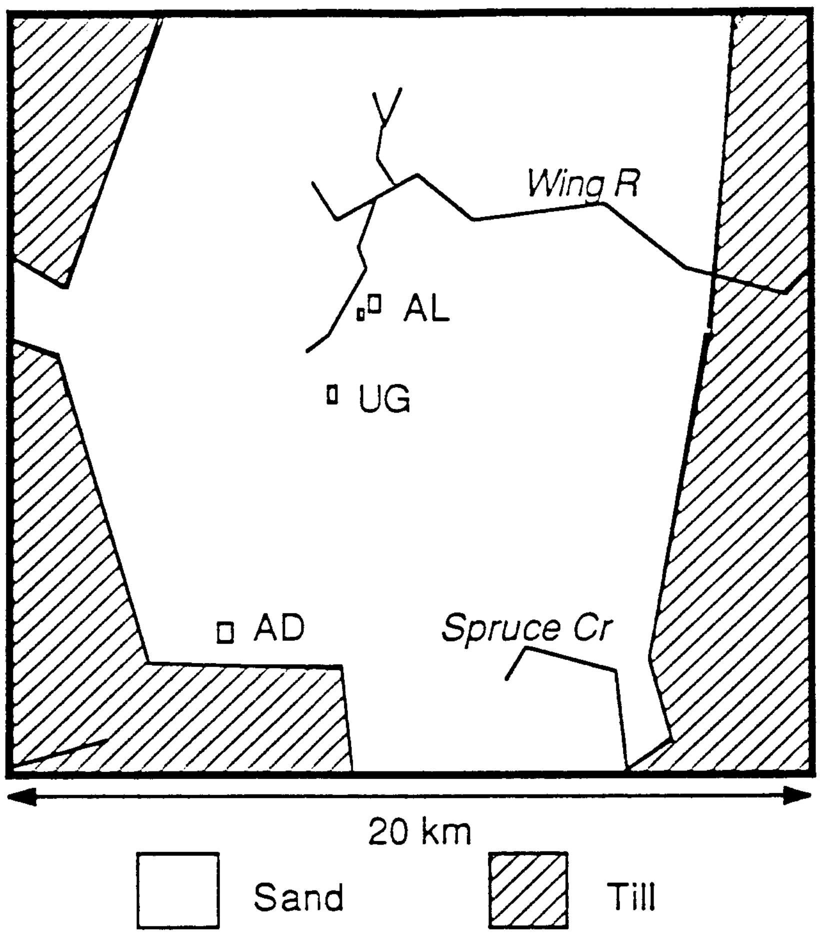

Fig. 2.7. Layout of the central portion of the full groundwater model, which extends far beyond the edges of this window. AL = Almora Lake, UG = Upper Graven Lake, and AD = Lake Adley. Lakes were modeled as simple rectangular elements capable of groundwater discharge or recharge over the area of the lake. Only the southernmost sub-basin of Lake Adley was modeled, because the northern sub-basins are too shallow to sustain a lake-level drop of more than a few meters.

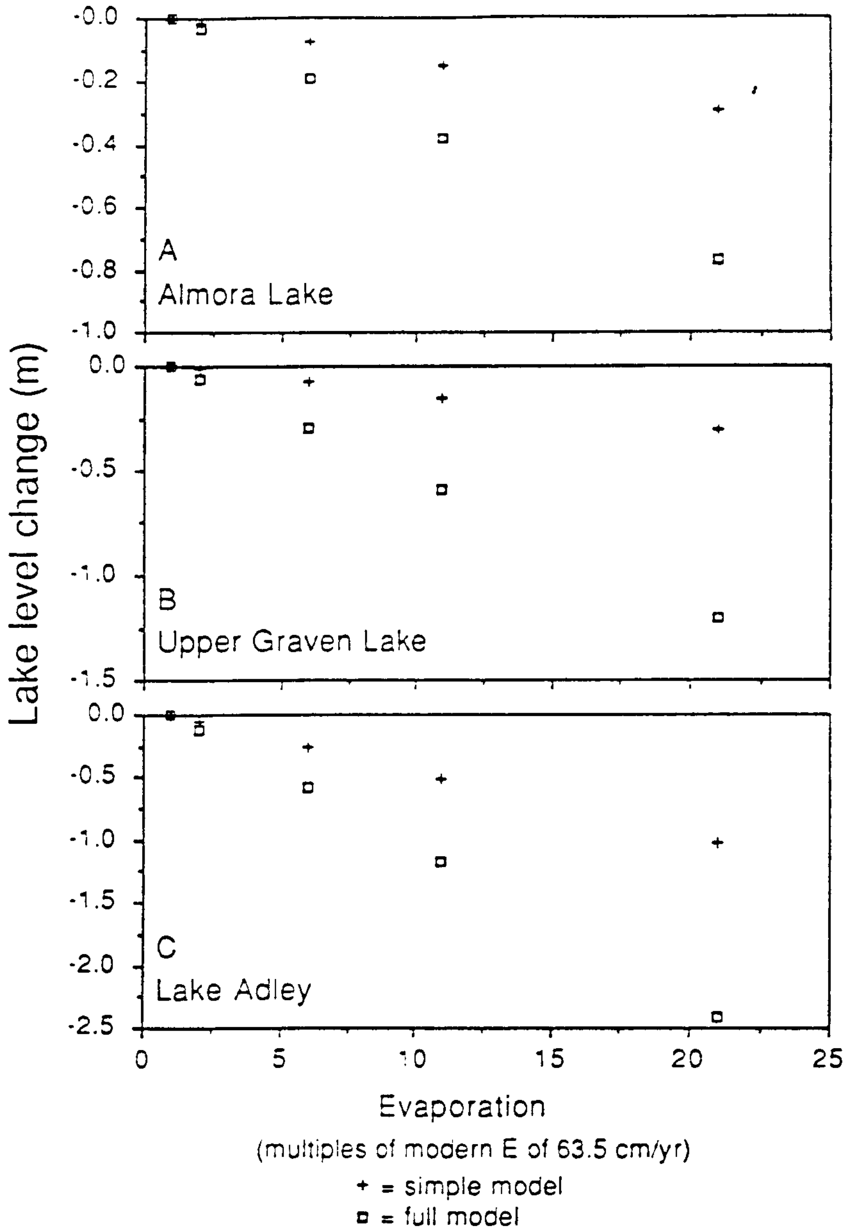

Fig. 2.8. Lake-level change as predicted by the simple and full groundwater models, for (A) Almora Lake, (B) Upper Graven Lake, and (C) Lake Adley in the Parkers Prairie sandplain. Note the different scales of the ordinates; Almora (closest to the river) is the least sensitive, and Adley (farthest from the river) is the most sensitive. Evaporation is plotted along the abscissa; the corresponding values of lake pumping g may be calculated as evaporation minus precipitation, assuming precipitation = 63.5 cm yr^{-1}. The extremely high hydraulic conductivity of the sandplain (2.25E-3 m s^{-1}, coarse sand and gravel) makes the lakes insensitive to the effect of γ (and hence also of evaporation), and unrealistically high values of evaporation were required to force lake levels to change on the order of a meter.

levels occurs within a distance of five or ten lake radii away from the river, and beyond that distance the river has little control on how lake pumping can change lake levels. (3) Lakes in highly permeable substrates are relatively insensitive to a change in the rate of lake pumping γ.

There are few implications of this study for modern systems. While there is an effect of lake position (proximity to a river) on lake-level sensitivity, the effect is somewhat weak and probably will not be particularly helpful in identifying sensitive lakes on the landscape. However, knowledge of the inverse proportionality between lake sensitivity and permeability may be useful.

For studies of paleo-lakes, knowledge of how lake pumping interacts with groundwater can be valuable in interpreting evidence of paleo-lake-level fluctuations in terms of past climate. For example, in order to investigate paleo-rates of E - P, lakes sensitive to those parameters must be chosen for study, namely large lakes not particularly close to a river. Conversely, large fluctuations in the level of a small closed-basin lake lying in highly permeable material cannot be explained in terms of E - P and must be attributed to another factor, probably a shift in the groundwater recharge over the region (Chapter 1).

In conclusion to both Chapters 1 and 2, in terms of lake-level change, the separation of the regional effect of shifting groundwater recharge from the local effect of shifting lake pumping is arbitrary. In reality, given the proper watershed, all of the lakes will be connected to each other via the water table, and changes in one lake level will have an effect, however small, on the levels of other lakes, and local lake-level changes will be superimposed over regional shifts in the elevation of the entire water table. The overall value of using a groundwater-based approach in studying past lake-level change is that, by choosing lakes that are sensitive to different factors, a particular configuration of the paleo-water table may be mapped, and hopefully that configuration will have been uniquely determined by a set of decipherable paleoclimatic conditions.

APPENDIX TO CHAPTER 2

Strack (1988) derives the equation of groundwater potential as a function of position for a discharge-recharge well couplet as depicted in Fig. 2A.1. This well configuration is the same as that presented in many textbooks as a simple example of a well and image well separated by a linear equipotential. For ϕ = piezometric head and k = hydraulic conductivity, the groundwater potential $\Phi = 0.5\, k\, \phi^2$. Then, following Strack, Φ as a function of position is given by

$$\Phi = [Q/(4\pi)] \ln\{[(x - d)^2 + y^2]/[(x + d)^2 + y^2]\} + \Phi_o \qquad (2A.1)$$

in which Q = pumping rate of each of the two wells, in units of m^3/s, and d = the distance of each well from the y axis (see Fig. 2A.1). In this system the y axis is an equipotential; other equipotentials are circles (as will be seen) eccentrically placed around each of the wells. To find the equation for one of these circular equipotentials, Φ is set to some constant value and then the variables of position are solved for. Setting $\Phi = \Phi_L$, and again following Strack (1988), eq. 2A.1 changes to

$$\Phi_L - \Phi_o = [Q/(4\pi)] \ln\{[(x - d)^2 + y^2]/[(x + d)^2 + y^2]\} \qquad (2A.2)$$

If the parameter α is defined in such a way that

$$-2\alpha = (\Phi_L - \Phi_o)\, 4\pi/Q \qquad (2A.3)$$

then eq. 2A.2 may be rewritten as

$$e^{-2\alpha}\, [(x + d)^2 + y^2] = [(x - d)^2 + y^2]$$

With some effort this equation may be rearranged in the following sequence:

$$x^2 - 2dx\, [(1 + e^{-2\alpha})/(1 - e^{-2\alpha})] + d^2 + y^2 = 0$$

$$x^2 - 2dx \coth(\alpha) + d^2 + y^2 = 0$$

and finally yielding

$$[x - d \coth(\alpha)]^2 + y^2 = [d/\sinh(\alpha)]^2 \qquad (2A.4)$$

From this last equation it may be seen that each equipotential is indeed a circle, centered at

$x_c = d \coth(\alpha)$ and $y_c = 0$, with a radius = $d/|\sinh(\alpha)|$.

Under Dupuit-Forchheimer conditions the circle mapped by eq. 2A.4 may be considered as the edge of a lake of radius r centered at a distance D from an infinitely long river of constant $\Phi = \Phi_o$. (In this case, no flow comes from infinity; the lake receives all its water from the river.) For the purposes of this paper it was imperative to convert eq. 2A.4 into a form containing the variables D, r, and Φ so that the sensitivity (*i.e.* the derivative) of Φ with respect to D or r could be calculated. From eq. 2A.4 and Fig. 2A.1 it is clear that

$$r = d/|\sinh(\alpha)| \quad \text{and} \quad D = d \coth(\alpha)$$

Solving both these equations for d, equating the results, and rearranging yields

$$D = r \cosh(\alpha) \qquad (2A.5)$$

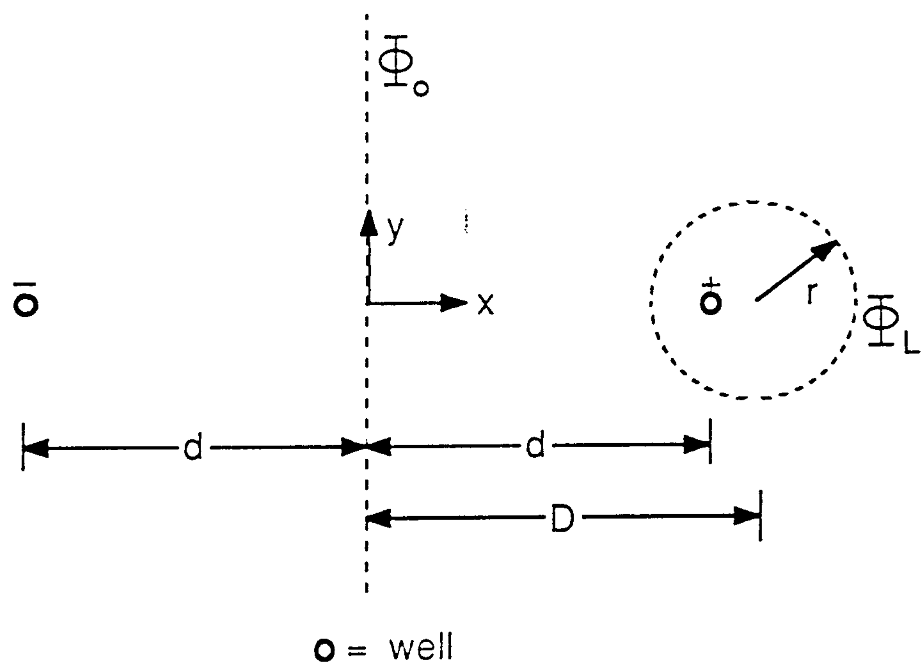

Fig. 2A.1. Schematic diagram showing discharge (+) and recharge (-) well couplet. The y axis is an equipotential at which $\Phi = \Phi_o$. Other equipotentials are circles eccentrically placed around the wells; one such equipotential is shown representing the edge of a circular lake of radius r centered at a distance D from the y axis.

This simple equation is the function relating D, r, and Φ for a circular lake next to an infinitely long river; however, Φ is embedded in α, and hence $\cosh(\alpha)$ must be decomposed:

$$D = r\,(e^{\alpha} + e^{-\alpha})/2$$

Rearranging and multiplying by e^{α}:

$$r\,e^{2\alpha} - 2D\,e^{\alpha} + r = 0$$

By the general solution for a quadratic equation, and some rearrangement:

$$e^{\alpha} = [D \pm (D^2 - r^2)^{0.5}]/r$$

In this case, the negative root is discarded because it does not give physically meaningful results. Taking the natural logarithm:

$$\alpha = \ln\{[D + (D^2 - r^2)^{0.5}]/r\}$$

But from eq. 2A.3

$$\alpha = (\Phi_o - \Phi_L)\,2\pi/Q$$

Equating these expressions for α and rearranging yields

$$\Phi_L = \Phi_o - [Q/(2\pi)]\,\ln\{[D + (D^2 - r^2)^{0.5}]/r\}$$

In this paper, the water is pumped from the lake not by a well but by net evaporation acting over the surface of the lake. Given a steady-state net evaporation (*i.e.* evaporation minus precipitation, or "lake pumping") of γ in units of m/s, then the discharge from the lake = $\gamma\pi r^2 = Q$. Substituting for Q results in

$$\Phi_L = \Phi_o - (\gamma r^2/2)\,\ln\{[D + (D^2 - r^2)^{0.5}]/r\} \qquad (2A.6)$$

which is identical to eq. 2.1 presented in the text (the subscript L is dropped in the text). This final equation satisfies the necessary condition that $\Phi_L = \Phi_o$ when r = D, *i.e.* when the lake is tangent to the river. Furthermore, as r approaches zero, the smaller lake area reduces the ability of evaporation to lower the lake below the river level; in other words, $\lim_{r\to 0} \Phi_L = \Phi_o$.

CHAPTER 3

LAKE AND GROUNDWATER HYDROLOGY DURING THE MID-HOLOCENE IN WEST-CENTRAL MINNESOTA

ABSTRACT

The response of a lake level to a shift in climate depends on characteristics of surficial and groundwater hydrology that are unique to that lake. Determination of the past levels of several lakes, rather than just one, should help provide a more unique determination of past hydrology and climate. Investigation of the sediments of several closed-basin lakes lying in the Parkers Prairie sandplain in west-central Minnesota indicates that lake levels were lowest about 8.5 to 8 ka (unpub. material; see Acknowledgments). I use a steady-state analytic-element groundwater model to quantify the effect of changing regional groundwater recharge N on the elevation of the water table in which these lakes reside. Because of limited overland flow on a sandplain, the surficial hydrology of the lakes can be simplified to the net quantity evaporation minus precipitation, which I label γ. I manipulate the N and γ of the model such that the modeled water table coincides with the paleo-lake levels for a given past time. Model results indicate that lake levels at 8.5 to 8 ka can be explained primarily by reducing N to 40% of the modern value, coupled with a γ of about 20 to 30 cm yr^{-1}. By 6 ka N had increased to 50 to 80% of the modern value allowing most lakes to rise in level, but γ may also have increased forcing at least one lake to remain nearly dry.

INTRODUCTION

Knowledge of past climates is important for understanding the causes of climatic change (Kutzbach, 1976) and for "providing vital verification and 'calibration' checks" of general circulation models (GCMs), which may help predict future global climate (Schneider and Dickinson, 1974, p. 489). For the late Pleistocene and Holocene epochs, much effort has been invested in inferring paleoclimate from fossil pollen spectra (*e.g.*, Webb and Bryson, 1972; Webb and Clark, 1977; Bartlein *et al.*, 1984). The use of pollen to predict past climate assumes, among other things, (1) that climatic change was the primary factor causing vegetation change, (2) that vegetation response to climatic change was rapid, and (3) that present relationships between climate and vegetation (or pollen spectra from surficial sediments) are truly applicable to the past. In general these assumptions are probably reasonable (Howe and Webb, 1983), and yet they are untested to the degree that an alternative method of inferring past climate over comparable temporal and spatial scales has been lacking. "Some independent measure of climate change is required to tighten up our understanding of vegetation-climate relations" (Ritchie, 1986, p 68).

One such independent approach is to use past lake levels to infer past climate. The relationship between lake levels and climate has been pondered for over 250 years (see Smith and Street-Perrott, 1983, for a brief review). It is generally recognized that the levels of closed-basin lakes (those without a surficial outlet) can be sensitive indicators of climate (Richardson, 1969; Street and Grove, 1979), and for the remainder of this paper I will be discussing *only* those lakes with closed basins. On a qualitative basis, high lake levels correspond to climates with high effective moisture, and low lake levels correspond to climates with low effective moisture. (The term "effective moisture" is intentionally vague and refers to some unspecified measure of precipitation inputs minus evaporation outputs from a lake and its catchment.) Qualitative changes in lake level have been shown to be geographically coherent over the last 30 ka (Street and Grove, 1979) and are therefore likely candidates from which to infer similarly qualitative changes in the effective moisture of past climatic regimes.

Quantifying the relationship between climate and lake levels permits not only the inference of past climate, but also the prediction of the effects on hydrology of future climatic change. This quantification has primarily been through a water-budget equation that models the surficial hydrological processes occurring in an endoreic (closed drainage) catchment. That is, the precipitation inputs are balanced against the evaporation and evapotranspiration outputs of the lake and its catchment (*e.g.*, see Street-Perrott and Harrison, 1985; Bradley, 1985, pp. 243-247). This approach has been used to indicate agreement between lake-level data and GCM model results for the past (Kutzbach and Street-Perrott, 1985). Kutzbach (1980) modified the approach into a combined water-and-energy budget by replacing the evaporation and evapotranspiration terms with expressions dependent upon the Bowen ratio (the ratio of sensible to latent heat loss from a surface) and the net radiation flux across the air-watershed surface.

In the quantitative inference of past climate from past levels of closed-basin lakes, simplifying assumptions have been made on at least two levels. First, the myriad of factors that influence how climate determines the regional hydrology of a catchment has been generally trimmed to just a few estimable quantities, such as precipitation, evaporation, and Bowen ratios. Second, the relationship between the regional hydrology of a catchment and the resulting lake level has also been greatly simplified. In particular, the effect of groundwater on lake levels has not been examined in paleoclimatic studies; indeed, some study sites have been chosen at least partly so that groundwater could realistically be ignored (*e.g.*, Kutzbach, 1980; Hastenrath and Kutzbach, 1983, 1985). Still, for many lakes groundwater cannot be ignored, leading Brakenridge (1978) to state "Because of the inseparability of lakes and groundwater, I am skeptical that any study which ignores the latter can yield valid paleoclimatic results." While my view is not quite so strict, a major purpose of this paper is to present examples of lake-level changes that are explicable primarily in terms of changes in the water-table elevation.

It is important to distinguish between lakes and the regional water table in which they lie. Perennial lakes are essentially expressions of the water table, which they modify via water-flux processes occurring at their surfaces, such as precipitation, evaporation, and receipt of overland runoff. All previous studies of which I am aware relating closed-basin lake levels and climate have assumed that lake levels result from a balancing of water inputs and outputs through the *local system* of the lake and its immediate catchment. In particular, to maintain a steady lake level, inputs must equal outputs, with high (gross) fluxes resulting in high lake levels, and low fluxes in low lake levels. It is true that changing the flux of water through a lake will change its steady-state level; however, it is not true that a change in lake level necessarily implies a change in water flux through the lake. Lake-level change could result from a change in the elevation of the *regional water table* that extends beyond the lake catchment, even with no appreciable change in water flux through the lake itself. Therein lies a danger in interpreting past lake-level changes solely in terms of a change in flux of lake water, and then inferring a paleoclimate from such calculated water fluxes.

Like an individual lake, the regional water table responds to a balance of inputs and outputs. Moreover, factors that tend to increase the flux of water through the groundwater system (and hence to raise the water table) often work in the same direction to increase the flux of water through a lake, hence producing a higher lake level than would have been caused by the rising water table alone. For example, an increase in precipitation P and a decrease in lake evaporation E and catchment evapotranspiration ET would tend not only to raise lake levels directly but also to increase groundwater recharge and thus raise the water table, further increasing lake levels. However, such factors (*e.g.*, P, E, ET, and recharge) do not always work in harmony; further, even when such factors do produce a qualitatively similar response of lake and water-table levels, the spatial pattern of the magnitude of the water-table response in general differs from the expected responses of individual lakes, where the effects of the local catchment hydrology are superposed independently on the regional water-table change. The water table responds to net inputs and outputs that do not occur at the same place: input from regional recharge (over a relatively large area) is balanced by outputs from other places, namely from river discharge (highly localized and linear) or from evaporative discharge from lakes and minerotrophic peatlands. Conversely, inputs and outputs of lake water occur at essentially the same place, namely over the surface or through the margins of the lake itself.

Lake levels, then, change in response to climatic and hydrologic forcing: (1) by being carried along with changes in a regional water table that extends beyond the lake catchment, and (2) by responding to a change in lake-water flux, which is superposed on the regional water-table change and modified by the groundwater hydraulics of the substrate surrounding the lake. Chapter 1 explores process (1) and describes the pattern of change to be expected in an interfluvial water table responding to a shift in regional recharge. Specifically, the water-table response is greatest for points farthest away from constraining rivers. Chapter 2 explores process (2), and begins by defining "lake pumping" as the net areal discharge (length time^{-1}) from the lake surface. Lake pumping, symbolized by γ, equals lake evaporation E minus the sum of direct precipitation P on the lake surface plus any inputs reaching the lake as overland runoff. For a region with minimal overland runoff, $\gamma \approx E - P$; note that γ will be negative if $P > E$. For a stable lake level, γ must be balanced by a net groundwater flux of equal magnitude but opposite sign such that the lake-water budget is balanced. Chapter 2 concludes that the sensitivity of a lake to γ is directly proportional to the radius of the lake and the distance from the lake to a constraining river and is inversely proportional to the permeability of the substrate in which the lake lies. Because lakes differ in their sensitivities to factors operating on different scales, *i.e.* local catchment hydrology versus regional groundwater hydrology, it is conceivable that the pattern of lake levels across a region integrated by a common water table would be a distinctive signature of the regional hydrology, and, by extension, of the regional climate.

In this chapter I present the results of applying this integration of groundwater and multiple lakes to the paleo-levels of a series of lakes sharing a common water table in the sandplain surrounding the village of Parkers Prairie in west-central Minnesota. I used an analytic-element groundwater model (Strack 1988) to quantify the effect of changing regional groundwater recharge on the elevation of the water table in which the lakes reside. Each modeled lake included "lake pumping" to quantify the net effect of local catchment hydrology on lake level. A group of researchers (see Acknowledgments) has been working on the stratigraphies of lake sediments from the region in order to determine the paleo-lake levels. From this work I have estimates of the levels of several lakes in the sandplain for two periods of time, about 8.5 to 8 ka and 6 to 5.5 ka. I then manipulated the regional groundwater recharge and local lake pumping in the model such that the modeled water table coincided with the paleo-lake levels indicated by the stratigraphic analyses. Lakes in the Parkers Prairie sandplain seem to have had their lowest levels during 8.5 to 8 ka, and to match those lake levels the model required about a 60% reduction in groundwater recharge (from the modern value of about 12.5 cm yr^{-1} to a paleo-value of about 5 cm yr^{-1}) combined with a lake pumping rate (E - P) of about 30 to 40 cm yr^{-1}. By 6 to 5.5 ka lake levels had risen, but the evidence quantifying this rise is not strong. Model results suggest that a 20 to 40% reduction in recharge (from modern values), coupled with a lake pumping rate of 50 to 80 cm yr^{-1}, could account for lake levels at that time.

These results probably have rather large errors that are difficult to estimate objectively. As more work is done on the stratigraphy of the lake sediment, estimates of paleo-lake levels will become more refined. For example, lake levels were probably even lower at 8 to 8.5 ka than those that I used in the model runs presented in this chapter (Björck, Feb. 1988, written comm.). Still, I emphasize that the pattern of lake-level evidence, and hence the model result, is much stronger at 8.5 to 8 ka than at 6 to 5.5 ka. The hydrologic change occurring between the two time periods should be viewed more as a qualitative trend than a shift from one quantified hydrologic regime to another.

This groundwater-based approach to studying lake-level change should aid our understanding of how lake levels are related to regional hydrology. However, I have left unquantified the relationship between regional hydrology and climate, at least as climate is usually presented in terms of average values of precipitation and temperature. Finally, while understanding the groundwater flow was important for interpreting the lake-level changes in the Parkers Prairie sandplain, there may be sites for which groundwater flow may be neglected with impunity. Large tropical or sub-tropical endoreic catchments for which the central lake basin is the lowest area in the regional water-table are good candidates (*e.g.*, Kutzbach, 1980; Hastenrath and Kutzbach, 1983, 1985).

STUDY SITE

The Parkers Prairie sandplain lies in the southeast corner of Otter Tail County in west-central Minnesota (Fig. 3.1). It occupies a saddle in the landscape, bordered by the higher Alexandria Moraine Complex and Henning Till Plain on the west and the Todd Drumlin Area on the east (Arneman *et al.*, 1969). North of the sandplain the landscape slopes rather gently toward the Leaf River, and south of the sandplain lies the relatively steep valley of Spruce Creek, draining southward to the Long Prairie River.

The surface of the sandplain grades to the north and east, implying that it was deposited by the Des Moines glacial lobe that occupied the Alexandria Moraine area to the west and the Spruce Creek valley to the south. The sand averages about 15 m in thickness, although it thins and feathers westward against the moraine and is over 30 m thick in the eastern portion of the sandplain (McBride, 1975). Kettle lakes are most common along the southwest border and central portions of the sandplain, where they occupy ice-contact deposits and an ancient drainage channel, respectively. The sandplain is drained primarily by the Wing River, which flows eastward before breaching a gap in the Todd Drumlin Area and turning north to join the Leaf River. The Wing River is surrounded by large fens (probably over 20 km^2); marl underlying some of this peat implies that the area was at one time an extensive lake, or several lakes. Modern lake levels across the sandplain, as well as groundwater modeling done by McBride (1975) and in my study, demonstrate that groundwater from about the southern third of the sandplain flows south toward Spruce Creek, despite the northward trend of the sandplain surface.

The region receives about 65 cm yr^{-1} of precipitation (Lindholm *et al.*, 1972, based on records from 1934-67), of which McBride (1975) estimated roughly 15 cm yr^{-1} recharges the water table in the sandplain; the remainder is presumably lost primarily as evapotranspiration, because overland runoff should be minimal across the sand surface. Lake evaporation might be similar to potential evapotranspiration (PET), estimated as about 58 cm yr^{-1} for Parkers Prairie from a statewide map of PET values presented by Baker *et al.* (1979), using the Thornthwaite method. However, assuming a pan coefficient of 0.7 and using a statewide map of pan evaporation values (Baker *et al.*, 1979), I calculated lake evaporation as being about 73 cm yr^{-1} for Parkers Prairie; national maps of shallow lake evaporation (Meyer, 1942; Linsley and Franzini, 1979) show a lake evaporation of roughly 76 cm yr^{-1} for west-central Minnesota. Considering the uncertainty in these figures, I concluded that lake evaporation was for all practical purposes the same as precipitation; Siegel and Winter (1980) also concluded that lake

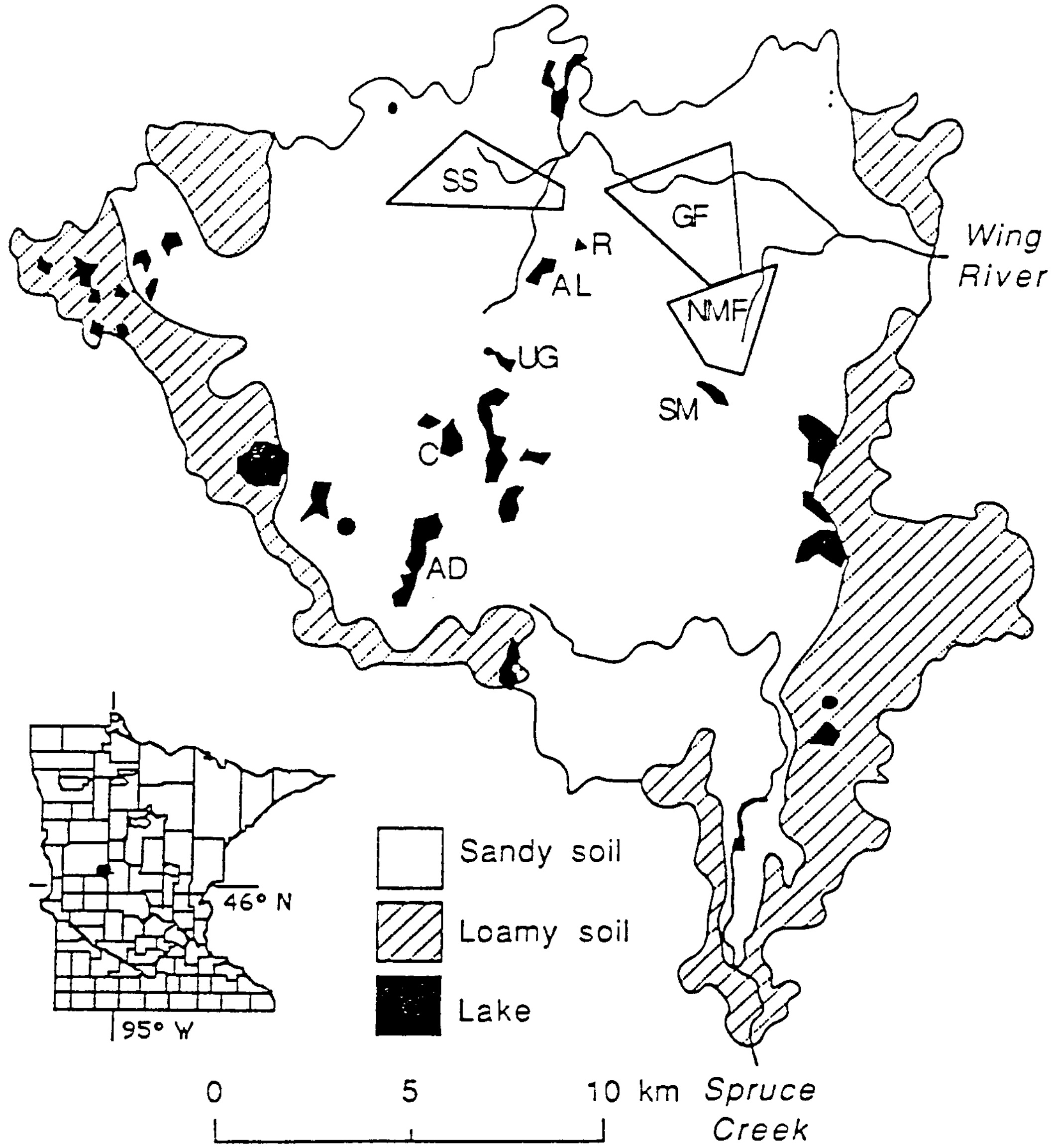

Fig. 3.1. Map of the Parkers Prairie sandplain area; the specific region shown includes the combined watersheds of the Wing River and Spruce Creek. Soil boundaries were taken from Arneman *et al.* (1969). R = Reidel Lake, AL = Almora Lake, UG = Upper Graven Lake, C = Cora Lake, AD = Lake Adley, and SM = South Maple Lake. The approximate boundaries of the peatlands overlying sand are indicated by the large quadrilaterals near the Wing River: SS = Simonson Swamp, GF = Griep Fen, and NMF = North Maple Fen. A similarly large peatland (not shown) lies near the head of Spruce Creek. The town of Parkers Prairie is situated just north of Lake Adley.

evaporation equaled precipitation, although their site lies about 85 km northeast of Parkers Prairie.

I chose the Parkers Prairie sandplain for many reasons. It lies near the triple-point of watershed divides for three major river systems (the upper Mississippi, Minnesota, and Red rivers), which seemed to be a good place to find changes in water-table and lake levels. Reconnaissance verified that lake levels apparently did fluctuate during the Holocene. Further, sandplains are regions for which many realistic assumptions may be made about both the surficial and groundwater hydrology, as will be explained later in greater detail.

EVIDENCE OF WATER-LEVEL CHANGES

Nature of the evidence

The term "lake level" has obvious meaning as an instantaneous measure of lake elevation. However, when viewed over any finite length of time, a "lake level" exhibits a variance and mean value that are to a degree arbitrary functions of the period of observation. To the paleohydrologist the "period of observation" is any window of time during which the lake leaves evidence of its level. The width of the window is determined by how fast the evidence can respond to lake level, and the window is smeared by sediment mixing and dating inaccuracies. Knowledge of the response times of different types of evidence would be valuable in interpreting the longevity and variance of paleo-lake levels.

The evidence of lake-level change is spatially discontinuous both locally and globally. Locally, lakes differ in the magnitude of their response to a given climatic change and in their ability to leave behind evidence of lake-level change. For example, a lake may have a critical depth so that strong evidence of lower lake levels may be recorded in its sediment, or perhaps a lake may lie in a substrate that is easily worked and modified by shoreline processes. Globally, closed-basin lakes are not distributed uniformly. They tend to be located in regions in which $E \geq P$, and they are often found in geologically young landscapes with poorly developed drainage systems, such as the glaciated regions of the Northern Hemisphere. However, even lakes that currently have open drainages (*i.e.*, have an outlet) may have been closed under drier paleoclimates (*e.g.*, Winkler, Swain, and Kutzbach, 1986) and cannot be discounted as sources of evidence for lake-level change.

The evidence of lake-level change is usually also temporally discontinuous. Whereas in some lakes there may be a relatively continuous record of qualitative lake levels (*i.e.*, whether the lake had "high" or "low" levels), most quantitative estimates for paleo-lake levels are restricted to discrete periods of time. To a degree temporal gaps in the record of quantitative lake levels are a function of methodology, *e.g.*, not enough sediment cores were taken, or the relevant information contained in a core is not interpretable. However, much of the lake-level record can be lost *because* of fluctuating lake levels. Specifically, extremely low lake levels may expose organic sediment to oxidation and allow higher shoreline features to be destroyed by erosion.

Finally, as noted above, not all lakes record the same climatic excursion with identical clarity. A lake may lie in a critical position in the watershed or be of a critical depth such that the lake sediments record some specific type of climatic shift that is undiscernible in the sediments of nearby lakes. Because the evidence of lake-level change tends to be discontinuous over both time and space, one should investigate as many lakes as possible in hopes that evidence from some lakes can fill the gaps in evidence from other lakes.

Types of evidence

Many different features of the biological, chemical, and physical components of lake sediments have been used to infer paleo-lake levels (see reviews by Richardson, 1969; Winter and Wright, 1977; Street-Perrott and Harrison, 1985). Water levels at Kirchner Marsh, Minnesota, were estimated with pollen and macrofossils (Watts and Winter, 1966), sedimentary pigments (Sanger and Gorham, 1972), and diatoms (Brugam, 1980). An interdisciplinary approach was likewise used at Lake Valencia, Venezuela (Bradbury *et al.*, 1981). The agreement among several independent types of stratigraphic evidence for lake-level change is important to consider, because factors other than lake level may contort the record of any one type of stratigraphic marker (Digerfeldt, 1986; Dearing and Foster, 1986). Despite the variety of stratigraphic techniques available to the paleohydrologist, most techniques give only a qualitative (or semi-quantitative at best) estimate of lake level. It is hoped that a series of these stratigraphic markers can be calibrated in the future, so that a single core from a deep portion of a lake could provide a continuous record of quantitative lake levels.

The most quantitative estimates of lake-level change have come from examining evidence of old shorelines. Subaerial strandlines demonstrating high paleo-lake levels have long been recognized in the more arid portions of the world, particularly in the southwestern United States and in central to north-central Africa. These higher shorelines have led to the identification of "pluvial" or "minevaporal" periods (Brakenridge, 1978; Street and Grove, 1979; Smith and Street-Perrott, 1983) and have allowed some quantification of past climatic parameters (Kutzbach, 1980; Hastenrath and Kutzbach, 1983, 1985). Paleo-lake levels lower than present may be found by searching for evidence of buried shorelines in sediment cores. Digerfeldt (1975, 1986) identified buried shorelines from a combination of sand and littoral plant macrofossils in the sediment column and then demonstrated that such paleo-shorelines could be reconstructed by following them in transects of cores from shallow to deep regions of the lake. Paleo-shoreline evidence is biased toward extreme lake stands, which may destroy evidence of intermediate lake stands.

Evidence of lake-level change in the Parkers Prairie sandplain

In order to fit a groundwater model to a past configuration of the water table, quantitative evidence of lake levels is needed from as many lakes as possible for a given time in the past. For the Parkers Prairie sandplain the strongest evidence comes from the time of lowest lake levels, about 8.5 to 8 ka; evidence from a few lakes shows that by 6 to 5.5 ka the water table had risen in some portions of the sandplain. These were the only two time periods for which I had sufficient lake-level information to justify the use of the groundwater model. The window of time over which each of these periods extends is unknown. As will be shown below, the low lake levels at about 8.5 to 8 ka persisted long enough to rework material at the lake margins into identifiable shoreline deposits. However, the higher lake levels at about 6 to 5.5 ka may have been only a brief transitional stand on the way to even higher levels. The 500 yr window given for each period is only a recognition of radiocarbon dating error due to old carbonate. I estimated this possible error as the difference in radiocarbon age between a piece of spruce wood and the limnic sediment surrounding it, collected from Upper Graven Lake; the sediment dated about 500 yrs older than the wood. Hence the uncorrected radiocarbon age of the lowest lake levels is 8.5 ka, but the true age may be closer to 8 ka; likewise the rising water table dated at 6 ka may actually be closer to 5.5 ka. It seems unwise to apply this one correction to all other radiocarbon ages in the sandplain, so I will present only uncorrected ages in the discussion that follows.

Almora Lake, Upper Graven Lake, and Lake Adley: In my study of the Parkers Prairie sandplain, lower lake levels were identified by two methods. The more accurate and straightforward method was that used by Digerfeldt, who, along with his colleague S. Björck of the University of Lund, Sweden, collected transects of sediment cores from Almora, Upper Graven, and Adley Lakes in the Parkers Prairie sandplain during March, 1986 (Fig. 3.1 and Table 3.1). In each lake they identified lower shorelines as sand layers containing a high concentration of *Scirpus* (bullrush) seeds. Knowing the growth habit of *Scirpus*, Digerfeldt estimated that paleo-lake levels were about 1 to 1.5 m above the sand and seed layer while it was being deposited. The three lakes had different magnitudes of lake-level lowerings, but the lowest lake levels were synchronous and occurred around 8.5 ka. The ages were obtained by Björck, who analyzed the pollen from the sand layer and its surrounding sediment. The pollen sequence through the sand layers corresponds to a shift from pine to prairie vegetation. This pollen transition has been identified and radiocarbon dated to about 8.5 ka in two long cores from the Parkers Prairie sandplain: Reidel Lake (analyzed by H. Jacobson, Univ. of Maine) and Upper Graven Lake itself (analyzed by S. Hobbie, Univ. of Minnesota). All of the work done in Parkers Prairie by Digerfeldt, Björck, Jacobson, and Hobbie is as yet unpublished and used only with their permission.

For 8.5 ka, then, Digerfeldt and Björck estimated lake-level lowerings of 2.5-3 m for Almora Lake and Upper Graven Lake, and 6-6.5 m for Lake Adley. There is little quantitative evidence for the levels of these three lakes at 6 ka, except that they were higher than at 8.5 ka. For modeling purposes at 6 ka, I subjectively chose lake levels based on my knowledge of the modeled water-table sensitivity (Table 3.2).

Cora Lake and South Maple Lake: The other method of identifying lower paleo-lake levels was simply to determine the radiocarbon ages of the basal limnic sediment of several shallow lakes in the sandplain, in this case Cora Lake and South Maple Lake (Fig. 3.1 and Table 3.1). If the lake-level lowerings estimated by Digerfeldt and Björck indicate a general lowering of the sandplain's water table at 8.5 ka, then some of the

Table 3.1. Locations and general information for the lakes and peatlands investigated; all sites lie in Otter Tail County, Minnesota.

Lake or peatland	Latitude, Longitude General Land Survey location	Area (ha)	Water depth (m)	Sediment thickness (m)
Reidel	46°12'43" N, 95°17'03" W SE1/4 SW1/4 sec. 25, T.132N, R.37W	4.0	3.82	15.83
Almora	46°12'21" N, 95°17'48" W NE1/4 sec. 35, T.132N, R.37W	16.5	ND	ND
Upper Graven	46°11'04" N, 95°18'25" W E1/2 SW1/4 sec. 2, T.131N, R.37W	6.4	1.11	5.30
Cora	46°09'54" N, 95°19'27" W NW1/4 NE1/4 sec. 15, T.131N, R.37W	29.3	2.27	2.67
Adley, S basin	46°07'42" N, 95°20'22" W SE1/4 sec. 28, T.131N, R.37W	ca. 17 (entire lake: 87.6)	6.05	>8.86
South Maple	46°10'29" N, 95°13'47" W E1/2 NE1/4 sec. 8, T.131N, R.36W	16.0	1.17	2.20
Simonson Swamp	46°13'24" N, 95°19'14" W NE1/4 NE 1/4 sec. 27, T.132N, R.37W	ca. 610	0	1.96
Griep Fen	46°13'11" N, 95°14'37" W NW1/4 SW1/4 sec. 29, T.132N, R.36W	ca. 790	0	2.85
North Maple Fen	46°11'57" N, 95°14'11" W NW1/4 SE1/4 sec. 32, T.132N, R.36W	ca. 460	0	4.66

Notes: ND = no data available; Almora was cored only along a near-shore transect. Lake location generally corresponds to the central portion of the basin, or sub-basin, that was cored; lake areas were determined by planimetry from topographic maps. Peatland location refers to the core site; while the sites were near the peatland middles, the sediment thickness may not be representative of the maximum thickness because of irregularities in the surface of the underlying substrate. Peatland areas are the areas of the four-sided polygons used to represent the peatlands in the groundwater model. Griep Fen and North Maple Fen are arbitrary subdivisions of one large peatland.

Table 3.2a. Summary of estimates of paleo-lake-level changes for the Parkers Prairie sandplain area for approximately 8.5 ka (uncorrected). Included are modeling errors and ranges for paleo-lake levels for the purpose of expressing relative confidence.

Lake Basin	Model[1] error (m)	Lake-level[2] change (m)	Constraining[3] range (m)	Confidence weighting values[4] Per basin	Sum for each lake
Reidel	-1.07	-2.75	14.5	0.064	0.064
Almora 1	-0.40	-2.75	0.5	1.324	2.647
Almora 2	-0.53	-2.75	0.5	1.324	
Upper Graven	-0.61	-3.0	1.5	0.882	0.882
Cora	-0.75	-4.4	0.5	2.647	2.647
Adley 1	+0.64	-6.25	0.5	0.662	2.594
Adley 2	+0.70	-6.25	0.5	0.662	
Adley 3	+0.73	-6.25	0.5	0.662	
Adley 4	+0.82	-6.25	0.5	0.608	
South Maple 1	+1.11	-2.75	0.5	0.901	1.969
South Maple 2	+0.94	-2.75	0.5	1.068	
Average absolute model error	0.76				

Notes:
1Model error = model elevation - lake elevation estimated from topographic maps, in m above MSL
2Lake-level change = lake level at 8.5 ka minus lake level at present
3Constraining range = subjective assessment of the reliability of the estimate of lake-level change; a wide range implies a low reliability. The measure is *not* a statistical measure of dispersion. See text for explanation.
4Confidence weighting value per basin = (1/greater model error) x (1/constraining range) / (number of basins in the lake). "Greater model error" is the absolute value of either the actual model error (column 2 above) or the average model error, whichever is greater. See text for explanation.

Table 3.2b. Summary of estimates of paleo-lake-level changes for the Parkers Prairie sandplain area for approximately 6 ka (uncorrected). Included are modeling errors and ranges for paleo-lake levels for the purpose of expressing relative confidence.

Lake Basin	Model[1] error (m)	Lake-level[2] change (m)	Constraining[3] range (m)	Confidence weighting values[4]	
				Per basin	Sum for each lake
Reidel	-1.07	-2.5	11.7	0.080	0.080
Almora 1	-0.40	-2.5	2.75	0.241	0.482
Almora 2	-0.53	-2.5	2.75	0.241	
Upper Graven	-0.61	-2.75	4.44	0.298	0.298
Cora	-0.75	-3.22	0.5	2.647	2.647
Adley 1	+0.64	-4	6.25	0.053	0.208
Adley 2	+0.70	-4	6.25	0.053	
Adley 3	+0.73	-4	6.25	0.053	
Adley 4	+0.82	-4	6.25	0.049	
South Maple 1	+1.11	-2.75	0.5	0.901	1.969
South Maple 2	+0.94	-2.75	0.5	1.068	
Average absolute model error	0.76				

Notes:
1Model error = model elevation - lake elevation estimated from topographic maps, in m above MSL
2Lake-level change = lake level at 6 ka minus lake level at present
3Constraining range = subjective assessment of the reliability of the estimate of lake-level change; a wide range implies a low reliability. The measure is *not* a statistical measure of dispersion. See text for explanation.
4Confidence weighting value per basin = (1/greater model error) x (1/constraining range) / (number of basins in the lake). "Greater model error" is the absolute value of either the actual model error (column 2 above) or the average model error, whichever is greater. See text for explanation.

shallower lakes in the sandplain should have been nearly if not completely dry, and hence not depositing sediment. The age of the basal limnic sediment would identify the time by which the water table had risen enough so that the lake basin held enough water perennially to deposit sediment. Presumably the basal dates from an array of shallow lakes that differ in depth or sensitivity could be used to develop a chronology of the rise in the sandplain's water table from its lowest position at 8.5 ka up to the present water-table elevation.

Care must be taken in the collection and interpretation of basal limnic sediment. Even if a basin is dry or too shallow to deposit much sediment, there may still be some residual organics left behind by marsh-like conditions or by earlier limnic conditions. These organics would give an anomalously old age for when the lake became perennially water-filled. Therefore I avoided any sediment mixed with the coarse sands and gravels that underlie the lakes, and for radiocarbon dating I chose the lowest fine-grained highly organic sediment immediately above the sands. The basal limnic sediment from Cora Lake dated to about 6 ka, and that from South Maple Lake to about 5 ka. The interpretation of the water depth under which such limnic sediment was deposited is problematic. Over several years I have seen the Parkers Prairie lakes fluctuate roughly a meter in depth, which is similar to annual fluctuations in other small lakes in Minnesota (Manson *et al.*, 1968) and Wisconsin (Novitzki and Devaul, 1978), and to annual water-table fluctuations in northern Wisconsin (Zaporozec, 1980). Assuming that a minimum depth of about 1 m is required to deposit organic limnic sediment, and assuming an annual lake-level fluctuation of 1 m (or $\pm$ 0.5 m), a minimum average lake depth might be 1.5 m while the basal limnic sediment was being deposited. Admittedly the lake level might have been higher and the water deeper than 1.5 m at that time; however, the water was appreciably shallower than 1.5 m just prior to the basal sediment age, else the lake would have been depositing sediment earlier. Hence as the lake level rose and the water depth increased from less than 1.5 m to equal to or greater than 1.5 m, at some time near the age of the basal limnic sediment the lake *was* 1.5 m deep.

Generally, little can be inferred about lake level prior to the onset of organic sedimentation, except that the water depth was probably shallower than about 1 to 1.5 m; indeed, the basin may have been completely dry and the water table many meters below the bottom of the lake basin. However, the complete drying of a basin should destroy any previously accumulated organic sediments. Both Cora Lake and South Maple Lake have a short segment of sediment (about 20 to 25 cm) that is transitional between the organic limnic sediment above that was radiocarbon dated and the inorganic pre-lacustrine sands and gravels below. Both of these core segments contain enough spruce and pine pollen (S. Hobbie, personal communication) to imply that at least part of the organics in these segments is residual from when a forest of spruce, and later pine, occupied the sandplain, about 12 to 8.5 ka. It seems unlikely that the water level in the basin ever dropped below this layer of residual organics, else they would have been destroyed by oxidation. The pollen grains in this layer were not degraded or corroded, and the sediment was not significantly humified. While I have no quantitative information on lake levels during the times of spruce and pine occupation, even these shallow basins had enough water to deposit some sediment. Further, the low water levels in Cora and South Maple were likely to have occurred at essentially the same time as the lowest water levels in the three lakes investigated by Digerfeldt and Björck, *i.e.* at 8.5 ka. Hence at this time Cora and South Maple probably had some water (else the residual organics would have been oxidized), but they must have had less than 1 to 1.5 m of water (else they would have deposited more sediment). Assuming again an annual lake-level fluctuation of about ± 0.5 m, I would guess that the two lakes each had a water depth of about 0.5 m at 8.5 ka. By 6 ka the level of Cora Lake had risen enough so that it had a water depth of about 1.5 m, as reasoned above, while South Maple Lake was still too shallow (≈0.5 m deep?) to deposit sediment (Table 3.2).

Reidel Lake, Simonson Swamp, Griep Fen, and North Maple Fen: Water-level data from these sites (Fig. 3.1 and Table 3.1) are essentially qualitative yet can aid the understanding of the hydrology of the Parkers Prairie sandplain during the early and mid-Holocene. The central sediment core from Reidel Lake gave no quantitative evidence of paleo-lake level, except that the basin has always been water-filled, as could be expected from the great depth of the mineral basin (about 20 m) and the insensitive position of the lake (close to the Wing River). For modeling purposes I assumed that Reidel Lake would have lake-level changes similar to those of nearby Almora Lake for both 8.5 and 6 ka (Table 3.2).

The history of the peatlands in the Wing River drainage is important because of their large area, which makes them sensitive to evaporation E and direct precipitation P; *i.e.*, a small change in E - P could have a large volumetric effect on the water balance of the sandplain. Unfortunately, peatland genesis and development is complicated and not well understood. Ideally the changing extent of the peatlands should be mapped over time by contouring a large collection of basal dates, bearing in mind the caveats about basal dates mentioned above. But such a mapping project was beyond the original scope of this project. Instead I collected one core from near the center of each of the major peatlands that I subjectively delimited from the topographic maps, and I used the basal age of each core to guide me whether to include that peatland as an evaporative surface in the 8.5 and 6 ka model runs.

Simonson Swamp is a forested peatland near the head of the Wing River (Fig. 3.1 and Table 3.1). The basal sediment is limnic and of unknown age: no pollen was found in preliminary raw smears. Despite the thin peat (196 cm), the site still may have been present as an evaporative surface in some form throughout the Holocene. Griep Fen and North Maple Fen (Fig. 3.1 and Table 3.1) have forested margins but in general are fairly open peatlands of sedge and cattail. The two fens coalesce to form the major continuous peatland in the Parkers Prairie sandplain. Spruce pollen at the base of the Griep Fen core and a spruce-needle trash layer at the base of the North Maple Fen core indicate that both these sites probably have been saturated surfaces (either open water or peatland) for the duration of the Holocene. North Maple Fen (and probably also most of Griep Fen) was a large shallow lake during the early to mid-Holocene, as indicated by 1.8 m of relatively clean marl at the base of the core. Thus, because of the old basal ages, I included all three peatlands in both 8.5 and 6 ka model runs. For lack of evidence I left the peatlands the same size as today, even though I suspect that they were significantly smaller in the past.

GROUNDWATER MODELING

Scope and assumptions of model

The groundwater model of the Parkers Prairie sandplain follows the segment of the hydrologic cycle beginning with the recharge of water across the surface of the water table and ending with the discharge of that water through surficial water bodies. The model ignores hydrologic processes occurring at the surface of the land, such as precipitation, evapotranspiration, overland flow, and interflow. For sandplains in general, the high infiltration rate and the gentle topography tend to make overland flow and interflow unlikely, and hence they could reasonably be excluded. Data from a study of a closed-basin lake on a sandplain in north-central Minnesota (Siegel and Winter, 1980) indicate that overland flow can be unimportant in such a setting: the rise in lake level following a major storm could be explained entirely in terms of precipitation and evaporation over the lake surface itself. Even during spring snowmelt coarse soils rarely form impermeable frost pavements that would allow significant overland flow (Storey, 1955; but see Schneider, 1961, for heavier soils). If one neglects overland flow, then recharge N to the water table would equal precipitation P minus evapotranspiration ET, and the surficial water budget of each lake could be collapsed into the lake-pumping term γ, which equals lake evaporation E minus P. The groundwater model can discern only N and γ, which are net quantities; the actual values of P, E, and ET remain unknown without an independent measure of one of them.

The groundwater model is a steady-state approximation of flow through the Parkers Prairie sandplain. Steady-state modeling smears all of the water-flux events (storms, snowmelt, droughts, etc.) over some period of time into one continuous average flux. Because of the large number of contributing factors, a change in the steady-state flux can be difficult to interpret. Nonetheless, the usefulness of the steady-state approach has been well accepted (Freeze and Cherry, 1979). It requires that the response time of the groundwater system be fast compared to the phenomena being investigated. For the regional, long-term paleoclimatic trends being investigated in this study, a steady-state approach seemed appropriate, especially given the fast response time of highly permeable systems such as sandplains. For example, McBride (1975) simulated irrigation pumping in his groundwater model of the Parkers Prairie sandplain, and the modeled water table reached a steady-state configuration after only a few years of pumping.

I used the analytic-element method of Strack (1988) to model the groundwater system. While some vertical flow may be discerned using principles of continuity (Strack, 1984; Haitjema, 1987), the model conforms to the Dupuit-Forchheimer assumption ($\partial\phi/\partial z = 0$, but also no resistance to flow in the vertical direction), and hence the model is essentially two-dimensional in the horizontal plane. The model allows for regions of different permeabilities, but the aquifer within each bounded region is presumed to be homogeneous and isotropic in the x-y directions. Likewise, regions of differing recharge may be delimited, but within each bounded region recharge is considered to be uniform. Lakes and peatlands were modeled simply as polygons over which recharge was set equal to the quantity P - E, which is the negative of lake pumping γ (and when E > P, the lake "recharge" becomes negative, *i.e.*, discharge to the atmosphere occurs).

The model is theoretically infinite in area, although in practice only its central portion is accurate. The consequence of the infinite-area approach is that a region very much larger than the area of interest must be modeled. For example, while the Parkers Prairie sandplain has a radius of about 7 to 8 km, I included in the model major rivers within a radius of up to 100 km; increasing detail was mapped toward the center of the model. The advantage of such a large modeled area is that groundwater divides fall out naturally on the modeled landscape. There is no need to surround the immediate area of interest with stringent boundary conditions, errors in which can cause large differences in the response characteristics of aquifers (Franke and Reilly, 1987). More importantly, such boundary conditions would have been largely unknown under the paleoclimates of interest in this study.

Model calibration and errors

"Model calibration" refers to the process of fitting the groundwater model to the present-day water-table configuration and groundwater flux over the region of interest. That is, the geometries of both the geology and the hydrology of the model must conform within reasonable limits to information estimated from field evidence or maps. Then, initial estimates of geologic and hydrologic parameters (*viz.*, permeabilities and water fluxes) are adjusted by trial and error until the water-table shape, groundwater recharge, and stream base flow in the model approximate known conditions.

The locations and elevations of rivers, lakes, and peatlands were taken from 1:24,000 topographic maps. The contact between the glacial deposits and the underlying bedrock was defined as the base of the aquifer and set at an elevation of 320 m (Lindholm *et al.*, 1972). The sandplain boundaries were estimated from soils maps (Arneman *et al.*, 1969). The thickness and permeability of the sand had been examined by McBride (1975), who had modeled the area in a U. S. Geological Survey study of the effects of irrigation on the sandplain. The sand averages roughly 15 m in thickness and overlies about 100 m of till; however, in the model I had to treat this entire thickness as being homogeneous. Because the total thickness of the aquifer is a tiny fraction of its horizontal extent (less than 1%), the water-table shape and groundwater flux are determined primarily by transmissivity, rather than by either aquifer thickness or hydraulic conductivity alone. (Transmissivity equals the product of the hydraulic conductivity and the saturated thickness of the aquifer.) Thus even though the modeled sand aquifer was much thicker than the actual sand aquifer, I was able to match

McBride's transmissivity values by lowering the model hydraulic conductivity relative to that estimated by McBride.

McBride quantified the flux of groundwater through the system by two methods. He estimated recharge by examining monitor well hydrographs, and then he estimated discharge by measuring the base flow of streams draining the sandplain; under steady-state conditions recharge should equal discharge. In the groundwater model I included lake pumping (E - P) from the lakes and peatlands, which was set equal to zero for purposes of model calibration, as mentioned before.

Given a certain stream configuration and aquifer thickness in a groundwater model, the topography of the water table is determined primarily by the ratio of recharge to hydraulic conductivity (N/K) within any bounded region uniform with respect to both N and K. The method of model calibration is to input N and K, and then examine the model output to check how closely the water-table elevations and streamflows match known values. I checked the modeled water-table elevations at the lakes investigated stratigraphically (see Fig. 3.1), as well as at a few other lakes in the till surrounding the sandplain. By trial and error I found the N/K ratios for the sandplain and till that seemed to give the best shape to the water-table. Most lakes in the sandplain were modeled to within 1 m of their elevation indicated on the topographic map ("Model error" column, Tables 3.2a and 3.2b). This error is reasonable when one considers that the lake levels currently can fluctuate a meter annually; moreover, elevations from U. S. Geological Survey topographic maps have associated errors of roughly one-half the contour interval (Robinson *et al.*, 1978), or 1.5 m for the maps of the Parkers Prairie area. The lakes in the till surrounding the sandplain were not fit nearly so well; while the qualitative shape of the water table was reasonable, the model elevation of the water table in the till was in error by 5 to 10 m at some points. Fortunately, the water table in the sandplain was not particularly sensitive to errors in the elevation of the water table in the till.

Once the water-table shape was fit, *i.e.* once the unique N/K ratio was determined, then N and K were manipulated in tandem (keeping N/K constant, or nearly so) to obtain a reasonable groundwater flux. McBride provided estimates for N, K, and the base flows (Q values) of streams draining the sandplain. Errors in these estimates are difficult to quantify, but Winter (1981) provides some guidelines. Values of K may be in error by at least 50 to 100%, not so much because of improper measurement techniques but because of the extrapolation of point data over the region of interest. Measurement of streamflow Q should be possible with an error <10%; however, one cannot be sure that the measured Q is representative of the base flow that corresponds to the presumed steady-state configuration of the water table. Winter does not discuss errors in the determination of recharge (N), but errors caused by the regionalization of point measurements of recharge should be similar to those errors in K values mentioned above. Still, the general uniformity of sandplains implies that regionalization errors would be lower than on other landforms.

Faced with such large unknown errors, I arbitrarily chose a range of ±20% around McBride's estimates of both N and Q, and I let K be determined by the N/K relationship defined by fitting the water-table shape. I then ran the model with different values of N (each with a corresponding K) to determine the maximum range in N that allowed values of *both* N and Q to be within 20% of those values estimated by McBride. That is, using both N and Q together should provide a better estimate of the groundwater flux than by using either N or Q alone. Throughout the analysis I used the Q value of the Wing River, although I did check the reasonableness of the Q values of the other streams draining the sandplain. I found that the lower estimate of groundwater flux corresponded to McBride's N value minus 20%; the higher estimate corresponded to McBride's Q_{Wing} value plus 20% (Table 3.3). I also determined a mid-range estimate of groundwater flux such that the percentage errors in N and Q_{Wing} (compared to McBride's values) were equal in magnitude but opposite in sign. For this run I also arbitrarily set $N_{till} = 0.1\ N_{sand}$, and allowed K_{till} to be whatever value provided a reasonable shape to the water table in the till regions surrounding the sandplain.

The above procedure provided a range of groundwater models, each one of which could be considered "calibrated" to the current water-table shape and groundwater flux: the extremes of this range are the low-flux model and the high-flux model (see Table 3.3). The question arises whether or not these two models have fundamentally different responses to a change in climate; if so, then each one would have to be examined separately to provide alternative hypotheses in explaining the causes of the lower lake levels indicated by the stratigraphic analyses. It is true that aquifers of lower K values are more sensitive to shifts in recharge N (Chapter 1). The lower K value of the low-flux model would therefore make it more sensitive than the high-flux model to a *unit* shift in N. However, the low and high-flux models have nearly identical sensitivities to a *proportional* shift in N: *i.e.*, a 40% drop in N would change the water-table configuration in essentially the same way in both the low and high-flux models. Thus only one model within the calibrated range needed to be fit to the paleo-lake levels, and the proportional shift in N required by that model would apply to the other "calibrated" models. I chose to work with the mid-flux model (Fig. 3.2 and Table 3.3) for examining model sensitivity and fitting the paleo-lake levels.

Model sensitivity

I examined the sensitivity of the mid-flux calibrated model with respect to recharge N, lake pumping γ, and elevation of the Wing River ϕ_{Wing}. That is, I performed a series of model runs in which all parameters were held constant except the one being examined. Knowledge of the way in which these parameters individually (although not necessarily independently) affect water-table and lake levels provides important insights into how these parameters might have been combined in the past to create the paleo-lake levels indicated by the sediment analyses.

The lake-level response to a lowering of N (Fig. 3.3) followed the pattern predicted in Chapter 1: lakes far from the Wing River dropped more than lakes close to the river. At 20% of modern N values, Lake Adley (far from the river) would be almost

Table 3.3. Parameters for the groundwater models considered to be calibrated to the modern water table.

Model	N_{sand} m s^{-1}	K_{sand} m s^{-1}	N/K_{sand}	N_{till} m s^{-1}	K_{till} m s^{-1}	N/K_{till}	γ cm yr^{-1}	Q_{Wing} m^3 s^{-1}	Q_{Spruce} m^3 s^{-1}
Low-flux	3.8E-9	1.50E-4	2.54E-5	7.1E-10	1.55E-6	4.58E-4	0	0.647	0.347
Mid-flux	4.0E-9	1.50E-4	2.67E-5	4.0E-10	8.00E-7	5.00E-4	0	0.616	0.322
High-flux	4.9E-9	1.67E-4	2.94E-5	1.0E-10	1.75E-7	5.71E-4	5	0.648	0.320

Notes: N = recharge, K = hydraulic conductivity, γ = E - P = lake pumping from lakes and peatlands, and Q = streamflow. Low-, Mid-, and High-flux refer to N_{sand}; a low N_{sand} allowed a high N_{till} with Q_{Wing} still remaining in the acceptable range, namely McBride's (1975) estimate of Q_{Wing} ±20%.

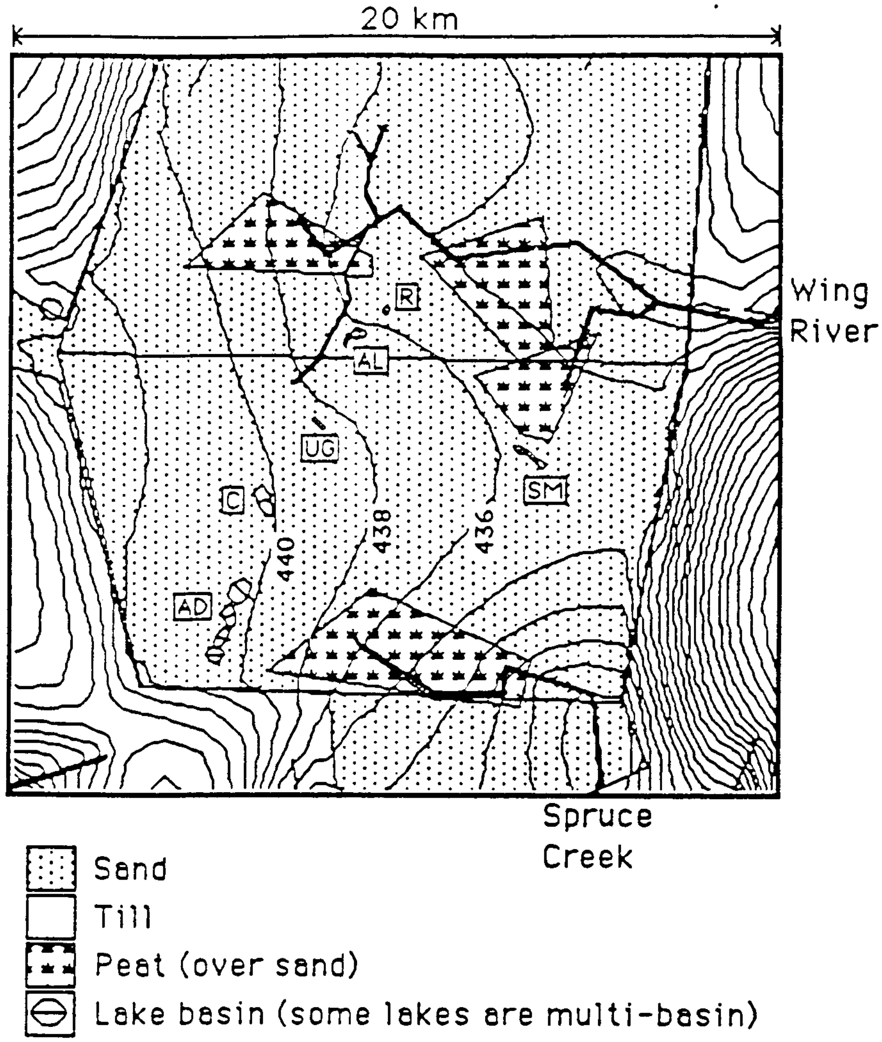

Fig. 3.2. Contour plot of the water table generated by the "calibrated" groundwater model using the "mid-flux" estimate of recharge (see Table 3.3). Elevation is in m above MSL; contour interval is 2 m. R = Reidel Lake, AL = Almora Lake, UG = Upper Graven Lake, C = Cora Lake, AD = Lake Adley, and SM = South Maple Lake. The bold lines are rivers; the thin horizontal line across the middle of the sandplain is just the connecting edge of several polygons placed over the sandplain to allow higher recharge there than on the adjacent till.

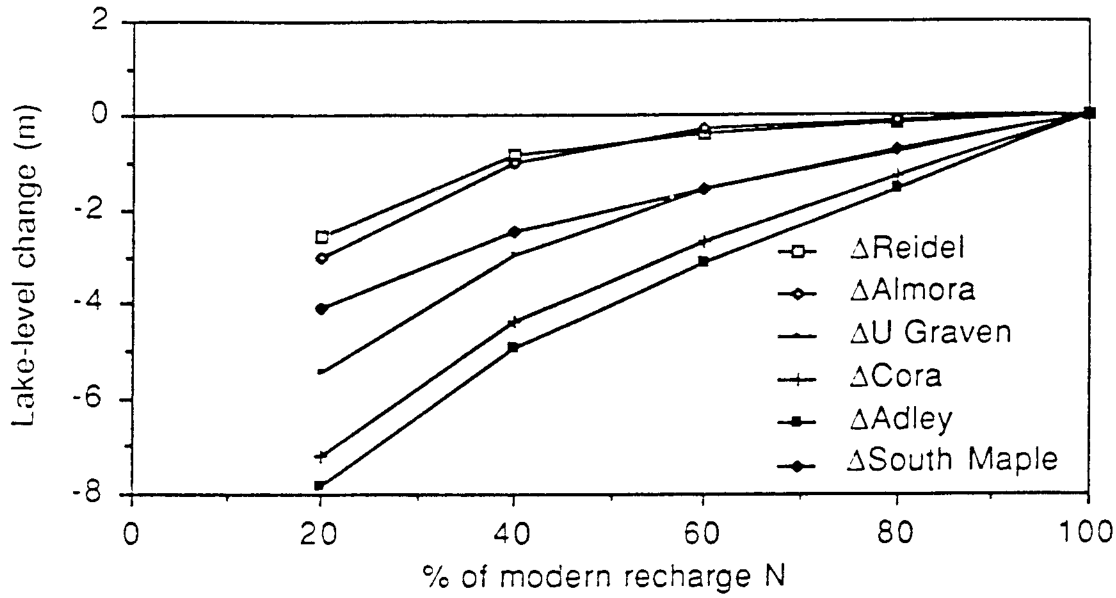

Fig. 3.3. Lake-level change as a function of the proportional change in recharge N. "100% of modern recharge" refers to the N of the mid-flux calibrated model, in which N = 4E-9 m s^{-1}, or 12.6 cm yr^{-1}. Throughout these runs, lake pumping and the Wing River elevation were held constant at the values used in the mid-flux calibrated model.

8 m lower than today, while Reidel Lake (close to the river) would be less than 2.5 m lower. The water table would have been roughly 4 m lower at South Maple Lake, which is lower than the bottom of the mineral basin: in other words, the lake would have been completely dry. The slopes of all the curves in Fig. 3.3 get steeper at lower values of N because the river is gradually drying up and getting shorter; as the river shortens the distances to the lakes increase, making the lakes more sensitive to a shift in N. Besides lowering lake levels, a drop in N also lowers the streamflow of the rivers draining the sandplain. As N drops from 100% down to 20% of the modern value, Q_{Wing} drops from about 0.61 down to 0.05 $m^3\ s^{-1}$.

Three points need to be made about the lake-level response to an increase in lake pumping γ (Fig. 3.4). First, the high hydraulic conductivity of the sandplain makes the lakes relatively insensitive to a shift in γ. Even with a γ of greater than 2 m yr^{-1} the most sensitive lake (Adley) dropped by only about 4 m, far short of the level indicated by the stratigraphic analyses for 8.5 ka. Further, most of the drop in lake levels was a regional result of γ acting on the large peatlands (see Figs. 3.1 and 3.2) near the Wing River -- not from γ acting on the lakes themselves. Other model results have shown that in the absence of the peatlands the direct local effect of γ on the lakes would lower them by only a few decimeters. Still, a lake may be sensitive to γ from a nearby peatland or large lake; *e.g.*, South Maple Lake is sensitive to γ from the adjacent North Maple Fen (Figs. 3.1 and 3.4).

Second, the lakes far from the Wing River (Adley and Cora) were in general more sensitive to γ than were lakes close to the river (Reidel and Almora), as might have been predicted by the results from Chapter 2. Again, however, I suspect that the evaporation from the peatlands rather than from the lakes themselves is the primary cause for this pattern. The large peatlands can evaporate enough water to cause a regional lowering of the water table, which would tend to lower more far from the rivers.

Third, the flattened part of the curve for South Maple Lake demonstrates a characteristic of the evaporative drying of lakes: if the surrounding water table is higher than the bottom of the lake basin, then evaporation cannot lower the water level in the lake (significantly) below the lake bottom. That is, evaporation may lower the water level right down to the lake bottom, but the constant inseepage of groundwater will maintain at least some degree of moistness -- perhaps enough to preserve some of the very basal organic sediments indefinitely. The lower bound on evaporative drawdown in lakes is in contrast to water-level lowering caused by reduced recharge N, in which case there is nothing to stop the water table from dropping below the bottom of the lake.

The final aspect of model sensitivity checked was the response of the lake levels to a drop in the elevation of the Wing River ϕ_{Wing} (Fig. 3.5). Under conditions of lower N or greater γ the flow of the Wing River must have been reduced, and the river might well have been somewhat shallower than today. However, the river is already

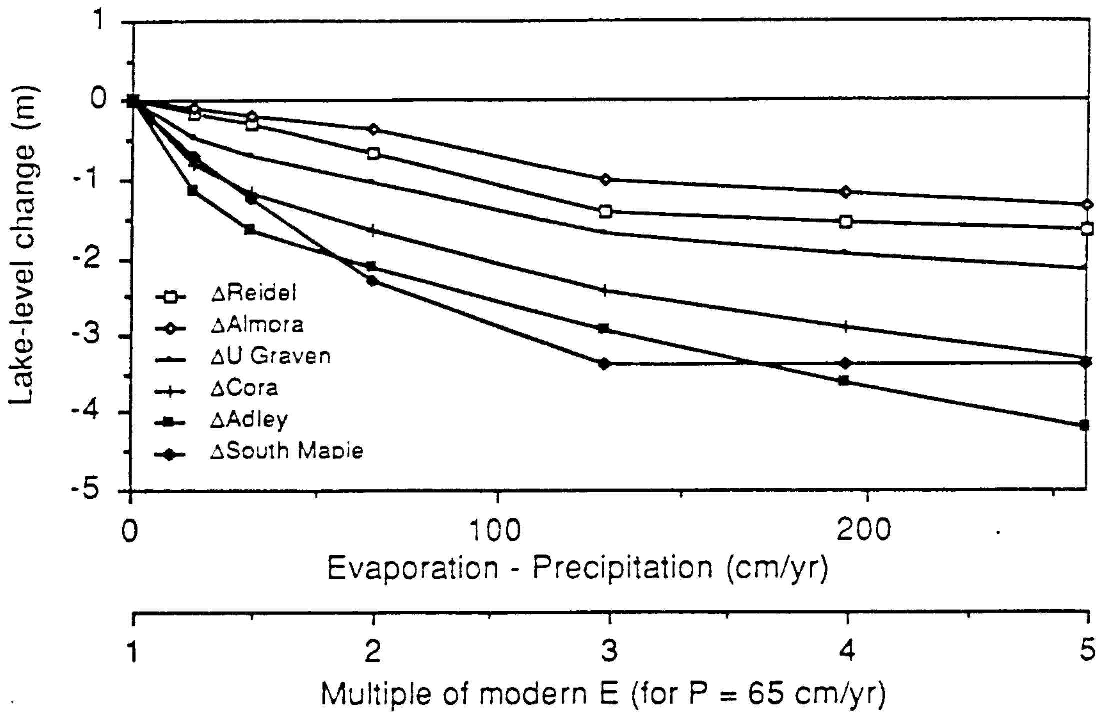

Fig. 3.4. Lake-level change as a function of lake pumping $\gamma = E - P$. The lower abscissa assumes that modern $E = 65$ cm yr^{-1}, and that P is held constant at 65 cm yr^{-1}. Throughout these runs, recharge and the Wing River elevation were held constant at the values used in the mid-flux calibrated model.

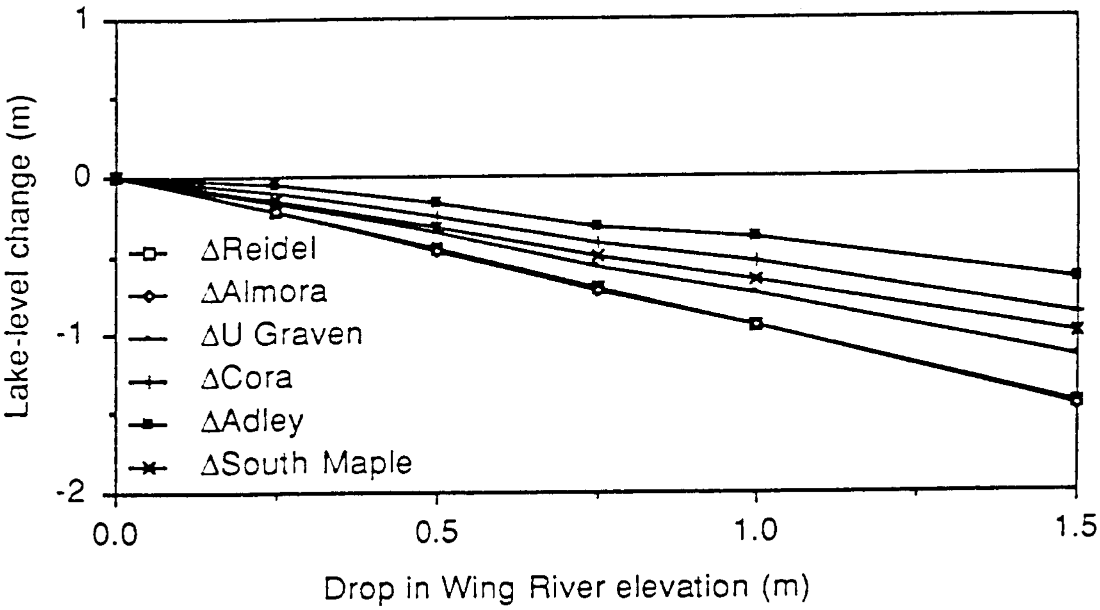

Fig. 3.5. Lake-level change resulting from a lowering of the Wing River. Throughout these runs, recharge and lake pumping were held constant at the values used in the mid-flux calibrated model.

rather shallow, and ϕ_{Wing} could not have been much lower in the past, perhaps a meter or so. Thus a lower ϕ_{Wing} could have only a minor effect on lake levels, and the model handles the possible effect in a rather naive manner. I merely lowered the entire river by a given amount, rather than attempting to calculate a new gradient for a given lower flow. While the magnitude of lake-level response is not great, the pattern is interesting because it is the opposite of that expected from a shift in either N or γ: lakes *close* to the river are obviously more sensitive to a change in river elevation than are lakes far from the river.

FITTING THE PALEO-WATER TABLE

The previous section outlining the sensitivity of the model to recharge N, lake pumping γ, and the elevation of the Wing River ϕ_{Wing} foreshadows the way in which these parameters must be combined in order to force the paleo-water table to fit the paleo-lake levels. In particular, lowering N is the only mechanism that is strong enough to account for the extremely low lake levels indicated by the stratigraphic evidence for 8.5 ka (compare Table 3.2a and Fig. 3.2, *e.g.* for Lake Adley). Increasing γ is too weak a mechanism on the sandplain to have been the sole cause of lower lake levels at 8.5 ka; however, it may have been involved to some degree. The goal of the modeling is to find which possible combinations of parameters (primarily N and γ) could adequately explain the paleo-lake levels. In other words, at 8.5 and 6 ka, how much of the lake level lowering was caused by reduced N, and how much by increased γ?

Fitting procedure

In the model low lake levels can be caused by lowering N, which drops the regional water table, or by increasing γ, which pumps water out of the lake via evaporation. Hence in theory a given low lake level could be matched in the model over a wide range of N values by simultaneously varying γ in such a way as to compensate for different N values. For example, presume a lake level could be matched at a certain N value and for $\gamma = 0$. If N is increased, the lake level could be kept low by pumping water out of it, *i.e.*, by allowing γ to increase. In general, then, a given lake level could be explained by a wide range of *paired* N and γ values. This range of values will be narrowed greatly by investigating as many lakes as possible: because each lake responds slightly differently to N and γ, including many lakes in the study will allow for a more nearly unique determination of N and γ.

To find an appropriate N and γ pair, I chose a certain value of N (*e.g.*, 50% of the modern value) and then manipulated γ until I obtained a subjectively determined best fit of the model to the paleo-lake levels. I then repeated this process with different N values to find other N and γ pairs, resulting in a range of N and γ pairs each of which could explain, to some degree, the lake-level evidence for a certain time in the past. Because of the non-exact match between the the present-day lake levels in calibrated model and those estimated from the topographic maps, the paleo-lake levels were fit based on the change in model lake levels rather than on the estimated elevation of the paleo-lakes.

Rivers were modeled as a series of ditches ("linesinks" in Strack, 1988), linked end-to-end in descending order of their elevation estimated from topographic maps. In the present-day calibrated model each of these segments is discharging, *i.e.*, the segment lies below the surrounding water table and thus pumps water out of the surficial aquifer. However, as N was decreased in the attempts to fit paleo-lake levels, the modeled water table dropped below the original input elevation of some of the upper river segments. In general I presumed that a river segment went dry if the water table dropped below it, and I would then remove that segment from the model. If the segment were not removed, it would become recharging, *i.e.*, the segment would pump water into the aquifer. It is extremely unlikely that any portion of the Wing River is recharging the sandplain because there is no steady-state source of water that could be shunted to the river to maintain its elevation higher than the surrounding water table. Overland flow is probably an insignificant source of water to the river on the sandplain for reasons noted earlier; further, because the sandplain is at the headwaters of the Wing River there is no upstream source of water.

However, a river should get shallow prior to going dry. I did not have an objective algorithm for changing the elevation and gradient of a river as the water table dropped. Instead, for lack of anything better I subjectively determined some rules to follow in lowering the Wing River. First, I assumed that the Wing River is no more than 1 m deep, at least over most of its length in the Parkers Prairie sandplain; in the model when the water table dropped more than 1 m below any river segment it went dry. Second, I noticed during the modeling runs that a reduction in recharge N always caused a proportionally larger reduction in the streamflow Q of the Wing River. For example, a 40% reduction in N would cause almost a 50% reduction in Q_{Wing}, if the elevation of the river segments were kept at the original input values and all recharging segments were removed. (As N lowers, groundwater divides shift on the landscape, and lower-elevation streams surrounding the sandplain capture proportionally more of the groundwater flow.) However, if the entire Wing River were lowered by 0.5 m, its new lower elevation would allow it to pump more water out of the sandplain and to recapture enough of its flow so that the total reduction in Q would now be only 45%. Thus my rule was to lower the Wing River enough, up to a maximum of 1 m, to recapture half of the flow reduction in excess of the proportional drop in recharge. That is, if a 40% reduction in recharge caused a 50% reduction in the flow of the unmodified Wing River, then I lowered the Wing River enough so that its flow was reduced by only 45%. To conclude, the lowering of the Wing River was subjective, small, and only affected those lakes quite near the river; nonetheless it was a useful exploration of lake-river interactions that provide a pattern of lake-level change that differs from the pattern caused by shifts in either recharge or lake pumping (contrast Fig. 3.5 vs. Figs. 3.3 and 3.4).

Index of Fit

Because of errors in the stratigraphic analyses and groundwater model, paleo-lake levels are not matched exactly by the model, and during each model run some modeled lake levels would come much closer than others to the estimated paleo-levels. I devised a subjective "index of fit" to evaluate how well all the paleo-lake levels were explained simultaneously by the water table of a particular model run. For each lake I calculated the difference between the lake-level change in the model and that estimated by the stratigraphic evidence. The index is essentially a sum over all lakes of the absolute values of these deviations; the better the fit, the smaller the index.

For several reasons it is not sensible to give each of these deviations equal weight. First, the quality of the groundwater model is spatially uneven, *i.e.*, the calibrated model is more accurate in some portions of the watershed than in others. Lakes lying in an accurate portion of the model should be given greater weight than other lakes. For each lake I calculated a "model error" as the difference between the calibrated model elevation and that estimated from the topographic maps (Table 3.2, column 2). Even though model accuracy should include the dynamics of the water table as well as its static configuration, for lack of better method I measured model accuracy as being inversely proportional to the calculated "model error." That is, for each lake I calculated a "modeling confidence" value as being the inverse of the absolute value of the model error (with exceptions noted below); I then used this modeling confidence to weight that lake's deviation calculated during the model runs in which I was attempting to fit the paleo-lake level. However, for reasons noted earlier the present-day lake levels can be estimated only with an error of about a meter or more. Further, a lake may be modeled "accurately" just because it happened to be fortuitously located at a point where the modeled water table was coincident with the estimated actual water table. To avoid giving a spuriously accurate lake an unduly high modeling confidence, I did not allow any lake to have a greater than average modeling confidence. That is, I averaged the absolute values of the model errors from all the lakes in the study; then for each lake I calculated the "modeling confidence" as the inverse of the absolute value of either the lake's actual model error or the average model error, whichever was greater (see notes, Table 3.2). In this way all the lakes modeled with reasonable (or greater than reasonable) accuracy were given the same average confidence weighting, while the inaccurately modeled lakes were given proportionately lower confidence weightings.

The other reason for not equally weighting the deviations of the modeled versus stratigraphically estimated lake-level changes is that the strength of the stratigraphic evidence differs greatly among lakes. The quality of the paleo-fit of the model should be determined primarily by those lakes for which the quantitative evidence of lake-level change is strongest. Other lakes whose levels are known only qualitatively for that same time period should be included in only a minor way in evaluating the model fit. For each estimated lake level for a given time in the past, I subjectively determined a "constraining range" (Table 3.2, column 4) that expressed my confidence in that estimate: the smaller the range, the higher my confidence. This constraining range was used only for weighting the relative importance of the lakes and is not a statistically derived quantity nor a definite range within which the paleo-lake level must have lain. For example, Digerfeldt and Björck (personal communication) estimated that at about 8.5 ka Almora Lake was between 2.5 and 3 m lower than today (Table 3.2a); the estimate of lake-level change I used was 2.75 m with a relatively small constraining range of 0.5 m, which reflects my high confidence in their estimate of paleo-lake level. However, compared to Almora, Upper Graven Lake is small and its shoreline features may not be so clearly or consistently defined. Thus I placed its 8.5 ka level at the low end of Digerfeldt and Björck's estimate (3 m) and arbitrarily increased the constraining range to 1.5 m to express my lesser confidence in this estimate. My estimate of the level of Reidel Lake at 8.5 ka is essentially only qualitative: while I can guess that its lake-level change was similar to that of nearby Almora Lake, all that I know for sure is that Reidel Lake dropped less than 14.5 m, which is the level at which the radiocarbon age of the sediment is about 8.5 ka. For each lake at each time in the past examined, I defined the "paleo-lake-level confidence" as the inverse of these subjectively-determined constraining ranges.

An overall confidence weighting for each lake basin for each past time period was then calculated as the product of the "modeling confidence" and the "paleo-lake-level confidence." Because for purely geometric reasons some lakes were modeled as a series of several basins (*viz.*, Almora, South Maple, and Adley lakes), the overall confidence weighting of any one basin was divided by the number of basins in that lake; thus the total confidence value for a lake (the sum of the confidences of the individual basins) was not an artifact of the number of basins. To summarize, for each basin

$$C_\Sigma = (C_M \times C_P)/B$$

in which C_Σ is the overall confidence weighting, C_M is the modeling confidence, C_P is the paleo-lake-level confidence, and B is the number of basins in that lake (see Table 3.2, column 5 and note 4). Lakes may be compared by examining the sum of the confidence weightings of their individual basins (Table 3.2, column 6).

As mentioned above, for each lake basin in each model run pertaining to a certain time in the past, I calculated a deviation between the lake-level change of the model versus that estimated stratigraphically. I then multiplied the absolute value of this deviation by C_Σ to get a weighted deviation for that basin; the "index of fit" value for a given model run was the sum of the weighted deviations of the all the basins for which I had paleo-lake-level evidence. That is,

$$I_F = (C_{\Sigma 1})\,|D_1| + (C_{\Sigma 2})\,|D_2| + \ldots + (C_{\Sigma n})\,|D_n|$$

in which I_F is the index of fit for a given model run, C_Σ is the overall confidence weighting of a basin, $|D|$ is the absolute value of the deviation of a basin, and the subscripts one through n refer to the basins having paleo-lake-level evidence. The I_F is a subjective value, particularly because of its inclusion of the paleo-lake-level confidence weightings. The primariy value of the I_F is as a precise way of explaining to the reader the subjective decision-making procedure I followed in trying to fit the groundwater model to the paleo-lake levels. Secondarily, the I_F may give some hint as to which model run -- *i.e.*, which pair of N and γ values -- might best explain the evidence of paleo-lake levels. Because each time period has its own specific set of paleo-lake-level confidences, the I_F values are comparable only among models of the same time period; *i.e.*, I_F values from the 8.5 ka model runs are not comparable to I_F values from the 6 ka model runs. The generally higher I_F values from the 8.5 ka model runs compared with those of the 6 ka model runs (see below) merely indicate that there are more lakes with strong enough lake-level evidence at 8.5 ka to contribute significantly to the sum that composes the final I_F value.

Results of fitting the model to paleo-lake levels

The lake levels at 8.5 ka could be fit, to a greater or lesser degree, by a range of models in which recharge N was reduced to 30 to 60% of the modern value and coupled respectively with a lake pumping γ of 0 to 60 cm yr^{-1} (Fig. 3.6). However, the best fitting model runs -- those with the lowest I_F values -- were those with N reduced to 40% of the modern value coupled with a γ of 20 to 30 cm yr^{-1}. Model runs with N greater than about 60% of the modern value could not match the low lake levels of Almora, Cora, and Adley without extremely high values of γ (greater than 60 cm yr^{-1}). A value of γ (*i.e.*, E - P) of 60 cm yr^{-1} currently corresponds to the extreme western portions of South Dakota and Nebraska; a γ of 75 cm yr^{-1} corresponds to southern Nebraska to eastern Kansas. The model runs with a γ of 20 to 30 cm yr^{-1}, a value that corresponds to portions of Manitoba, the eastern Dakotas, and southwest Minnesota, seem more reasonable. The specific model run in which N = 40% of the modern value and γ = 20 cm yr^{-1} is plotted in plan view (Fig. 3.7; compare to Fig. 3.2) and cross section (Fig. 3.8) to give a better graphical representation of what might have been the general water-table configuration across the sandplain at 8.5 ka.

The lake levels at 6 ka were fit by range of models with N reduced to 50 to 80% of the modern value and coupled with a γ of 30 to 80 cm yr^{-1} respectively (Fig. 3.9); the best fit was that with an N = 80% of the modern value with γ = 75 cm yr^{-1}. As mentioned above, such a value of γ seems high and probably should be viewed with skepticism for two reasons. First, in contrast to 8.5 ka when least five lakes show good quantitative evidence of lake-level change, the 6 ka model runs depended almost entirely on only two lakes, Cora and South Maple (compare lake-total confidence weighting values, column 6, Table 3.2b). Second, South Maple Lake was not fit particularly well by the model. During the calibration run, South Maple Lake had one of the highest "model errors" (Table 3.2, column 2) of any lake; during the 8.5 ka model runs, the lake-level change at South Maple was consistently over-estimated at about -3.9 m, instead of the targeted -2.75 m (Fig. 3.6g). Notwithstanding these concerns, the plan view (Fig. 3.10) and cross section (Fig. 3.11) of the water table for the specific model run in which N = 80% of the modern value and γ = 75 cm yr^{-1} pictorially represent what the water table configuration might have been at about 6 ka.

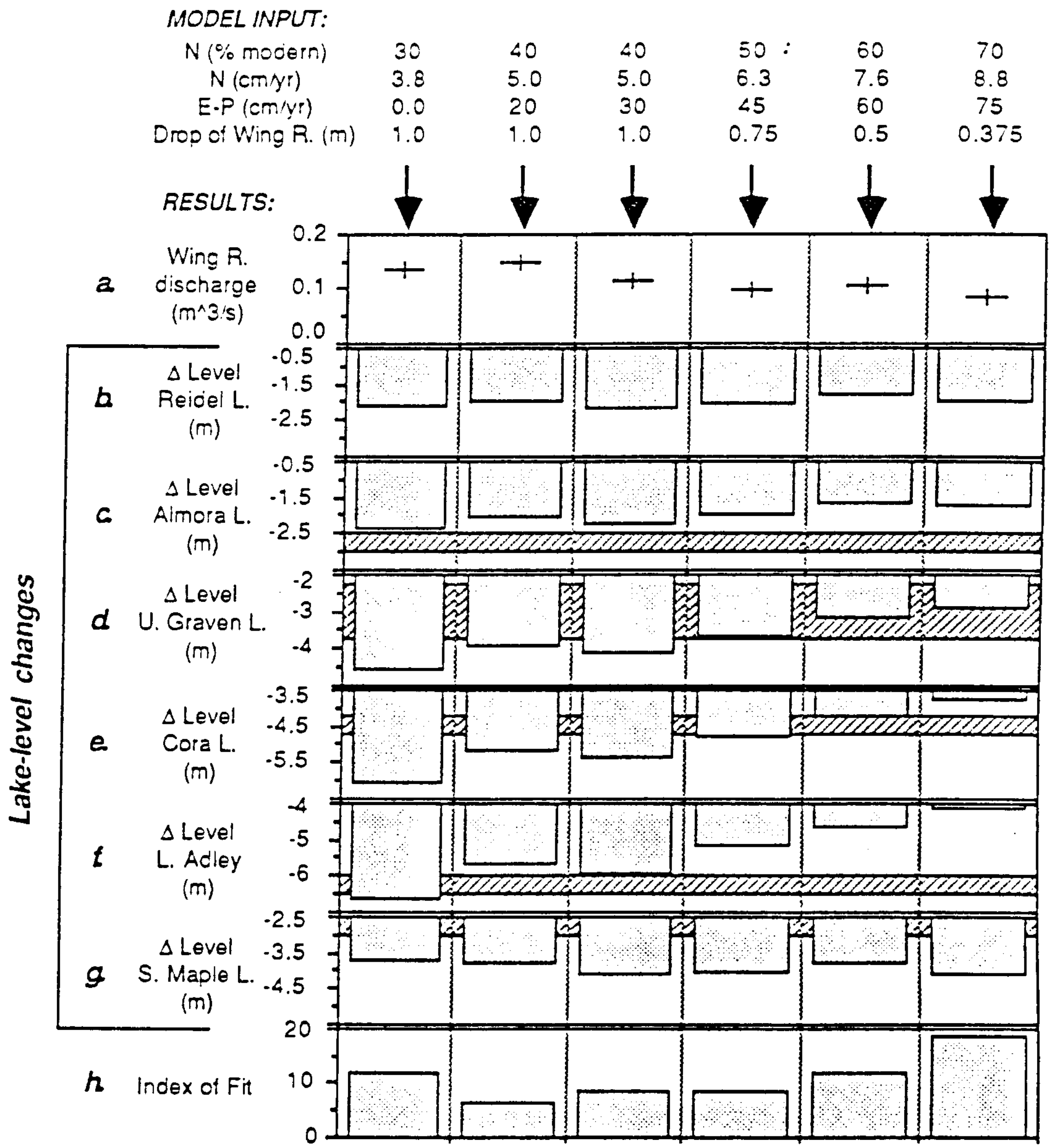

Fig. 3.6. Series of plots comparing the results of six model runs (listed across the top) attempting to fit the lake-levels at 8.5 ka.

a. Discharge from the Wing River is roughly 0.1 $m^3\ s^{-1}$, about one-sixth the modern value.

b. through *g.* Lake-level changes for each lake for each model run. The stippled bar diagram for each lake points downward; *i.e.*, a low lake level is indicated by a negative lake-level change. Note the different ordinate values: while each plot covers a 3-m range of lake-level change, some lakes are much lower than others. Cross-hatched band across some plots indicates the paleo-level for that lake as estimated by stratigraphic analyses. A "good fit" occurs when the lake-level change is near or within the cross-hatched band.

h. The "index of fit" for each model run, as explained in the text. The "best fitting" model runs have the smallest "index of fit" values.

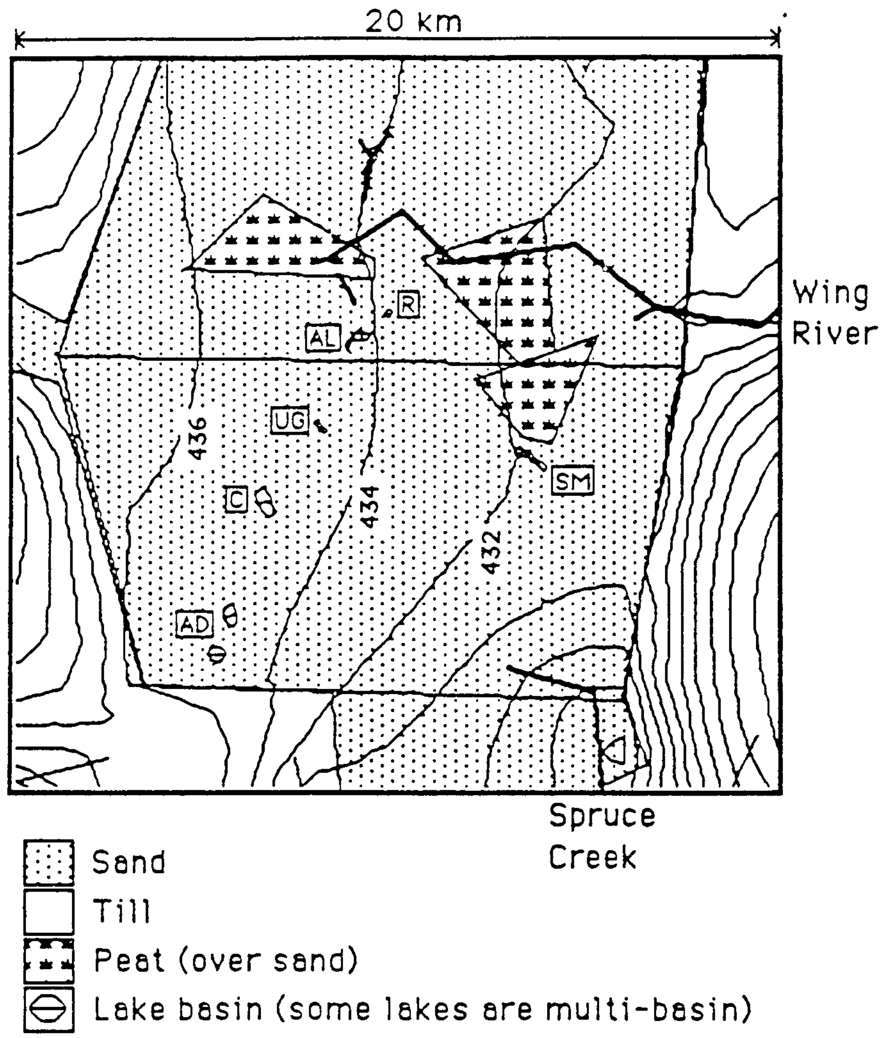

Fig. 3.7. Contour plot of the water table generated by the groundwater model attempting to fit the lake levels at 8.5 ka. In this model run recharge N = 40% of the modern value, E - P = 20 cm yr^{-1}, and the drop of the Wing River = 1 m; this model configuration had the lowest "index of fit" value (see Fig. 3.6, second "column"). Elevation is in m above MSL; contour interval is 2 m. R = Reidel Lake, AL = Almora Lake, UG = Upper Graven Lake, C = Cora Lake, AD = Lake Adley, and SM = South Maple Lake.

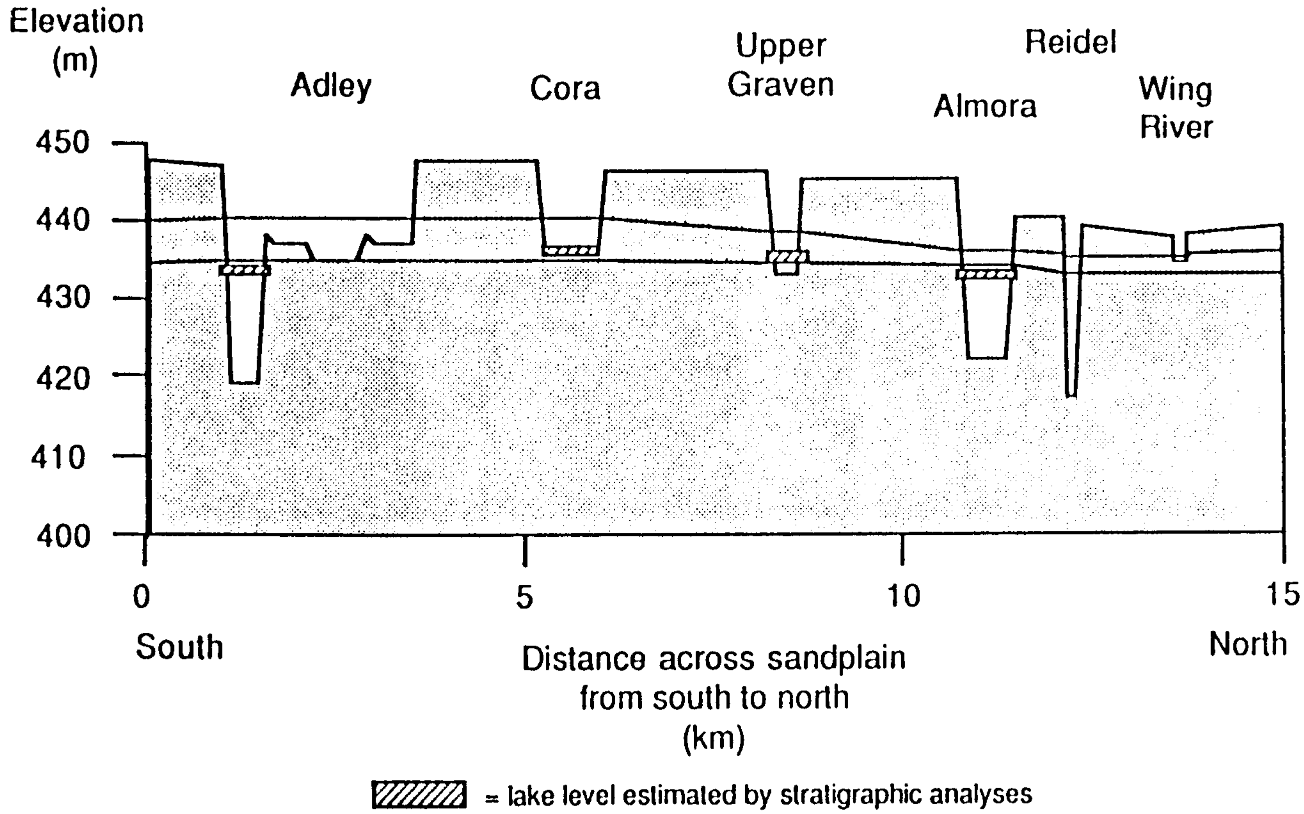

Fig. 3.8. Schematic cross-section through the sandplain showing:
(1: upper water table) the modern water table as estimated by the model;
(2: lower water table) the water table of the model run shown in Fig. 3.7; and
(3: crossed-hatched bars) stratigraphic estimates for lake levels at 8.5 ka.

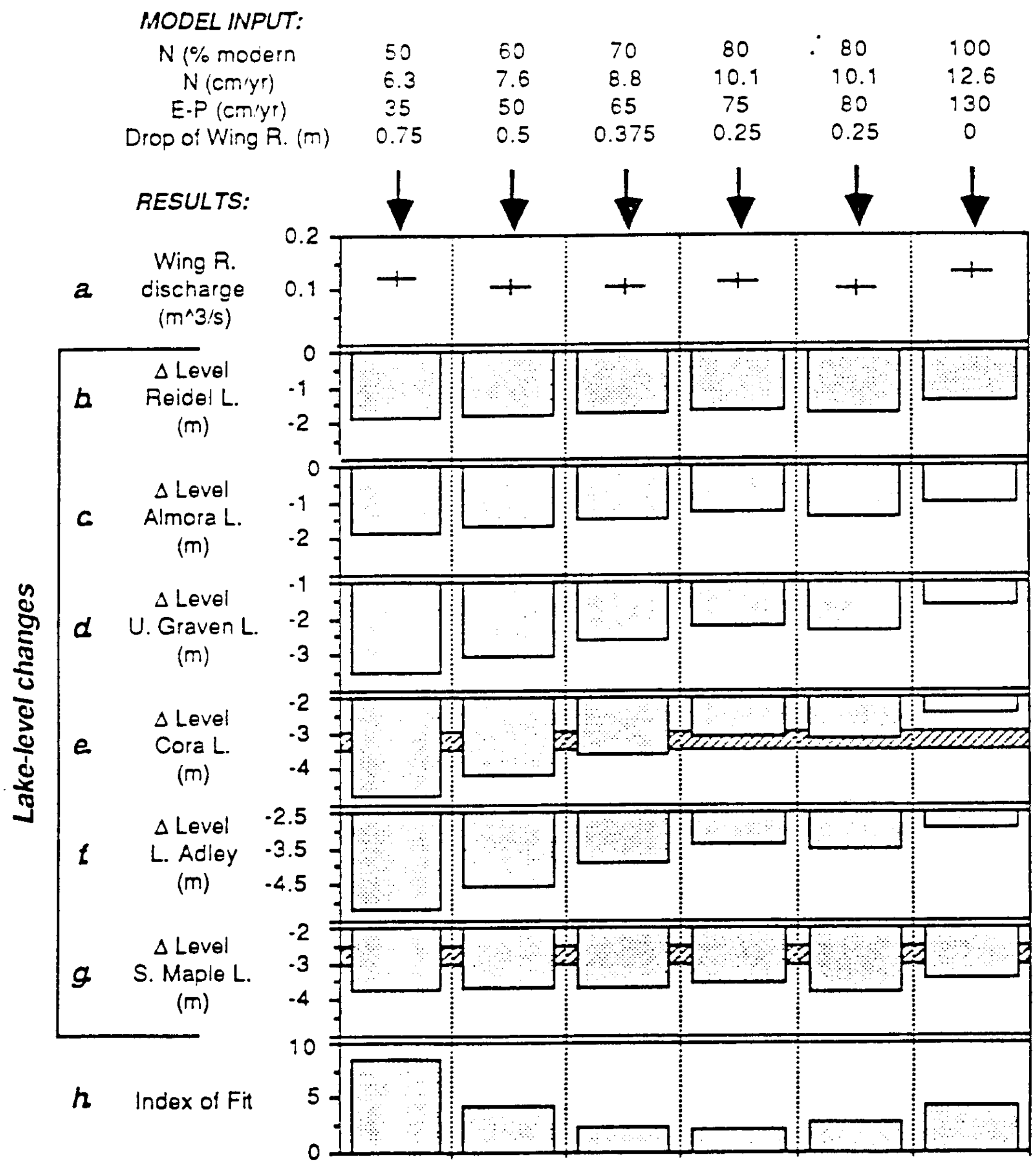

Fig. 3.9. Series of plots comparing the results of six model runs (listed across the top) attempting to fit the lake-levels at 6 ka.

a. Discharge from the Wing River is roughly 0.1 $m^3 s^{-1}$, about one-sixth the modern value.

b. through *g.* Lake-level changes for each lake for each model run. The stippled bar diagram for each lake points downward; *i.e.*, a low lake level is indicated by a negative lake-level change. Note the different ordinate values: while each plot covers a 3-m range of lake-level change, some lakes are much lower than others. Cross-hatched band across some plots indicates the paleo-level for that lake as estimated by stratigraphic analyses. A "good fit" occurs when the lake-level change is near or within the cross-hatched band.

h. The "index of fit" for each model run, as explained in the text. The "best fitting" model runs have the smallest "index of fit" values.

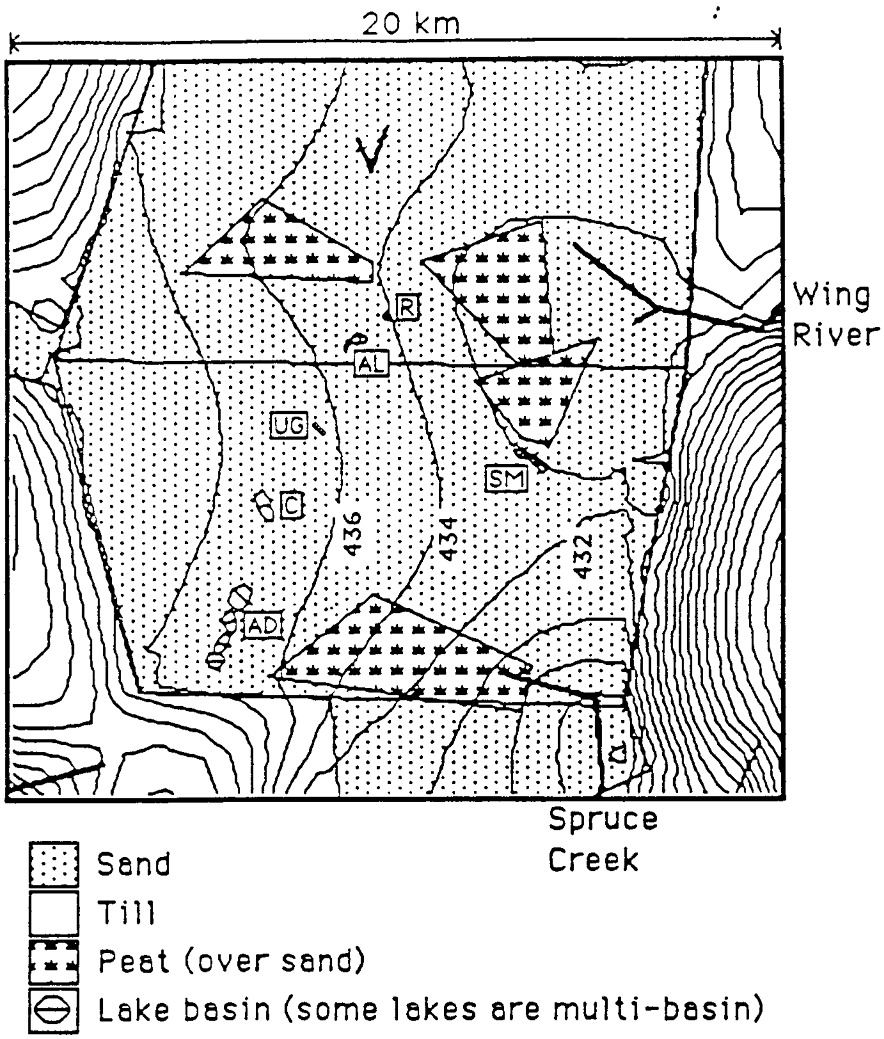

Fig. 3.10. Contour plot of the water table generated by the groundwater model attempting to fit the lake levels at 6 ka. In this model run recharge N = 80% of the modern value, E - P = 75 cm yr^{-1}, and the drop of the Wing River = 0.25 m; this model configuration had the lowest "index of fit" value (see Fig. 3.9, fourth "column"). Elevation is in m above MSL; contour interval is 2 m. R = Reidel Lake, AL = Almora Lake, UG = Upper Graven Lake, C = Cora Lake, AD = Lake Adley, and SM = South Maple Lake.

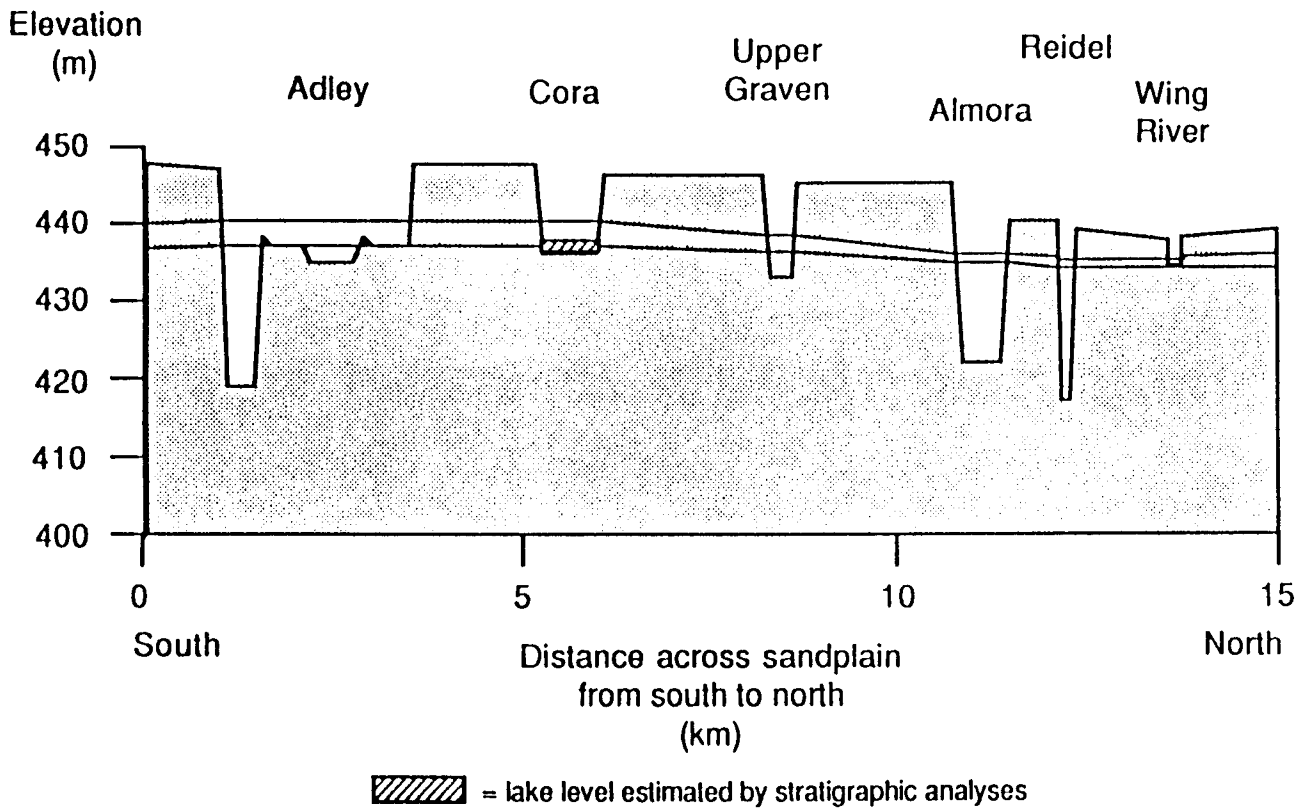

Fig. 3.11. Schematic cross-section through the sandplain showing:
(1: upper water table) the modern water table as estimated by the model;
(2: lower water table) the water table of the model run shown in Fig. 3.10; and
(3: crossed-hatched bar) stratigraphic estimate for level of Cora Lake at 6 ka.

Still, the trend in lake levels is still interesting even if the 6 ka-model parameters cannot be estimated with precision: at 8.5 ka the levels of both Cora and South Maple lakes were approximately at their bases (*i.e.*, the bottom of their mineral basins), and by 6 ka Cora had risen enough to begin deposition of organic sediment while South Maple was forced to remain at its base. The difference between the sensitivities of the two lakes helps explain this trend. Cora is more sensitive than South Maple to N (Fig. 3.3), while South Maple is more sensitive than Cora to γ (Fig. 3.4), particularly at higher values of γ. The rise in Cora during this time necessitates an increase in N, which in turn necessitates an increase in γ to keep South Maple at its base. The major unknown in this scenario is how events in the nearby peatland may have been affecting the water level in South Maple Lake. There exists the possibility that peat growth in the basin to the north of South Maple Lake may have been a more proximal factor than climate alone in raising the water table; however, the mechanism by which peat growth can be the cause, and not effect, of a higher water table is unclear and not fully explored in the literature.

DISCUSSION

Relation between model results and climate

For a given time in the past, the groundwater model results in a pair of N (recharge) and γ (lake pumping) values that best fit the model water table to the paleo-lake levels. Climate, on the other hand, is usually spoken of in terms of parameters such as average annual precipitation P and temperature T; indeed, general circulation models (GCMs) of the climate can produce output including not only P and T but also barometric pressures, wind vectors, humidities, and so forth. As mentioned earlier, both N and γ may be considered *net* quantities, which obscures their relationships to the component climatic parameters. Nonetheless it is important to investigate these relationships to better interpret the possible climatic meaning of a given groundwater model result, and to discern the ways in which the original groundwater model could be expanded and improved so as to mechanistically convert N and γ into more conventional climatic parameters.

On a sandplain with no significant overland flow, γ simply equals the quantity E minus P, with the obvious implication that an increase in γ can result from either an increase in E or a decrease in P. Further, an increase in E may well result from an increase in T; however, Hastenrath and Kutzbach (1983) point out that E and T are not simply related. On landforms other than sandplains, γ may have to include input from the catchment as overland flow, which is itself sensitive to P, ET, soil characteristics, and topography and would require a surficial hydrology model for quantification.

Again, if overland flow is not significant on a sandplain, then N simply equals the quantity P minus ET; hence either an increase in P or a decrease in ET would cause an increase in N. On other landforms overland flow may have to be quantified with a surficial hydrology model handling topography and soil characteristics as well as P and ET. While ET may ultimately be a function of climate, a change in vegetation can also shift ET. The control of ET by vegetation should be particularly evident along boundaries of major physiognomic groups of vegetation, such as the prairie/forest border, or even the coniferous/deciduous forest border. As a sweeping generality, under the same climate annual rates of ET are greatest for conifer forests, less for deciduous forests, and least for grasslands and scrublands (Bosch and Hewlett, 1982). A reduction in ET caused by forest clearance may in fact raise the water table (Moore, 1975; Peck and Williamson, 1987), presumably by increasing N. In north temperate states this increase in N may be the result of deforestation reducing ET during the critical spring snowmelt period, when most of the annual recharge takes place (Bowes *et al.*, 1984; Verry *et al.*, 1983). Moreover, annual values of P and ET are typically much larger than N: in the Parkers Prairie sandplain P is approximately 65 cm yr^{-1} while N is only about 12.6 cm yr^{-1}, leaving an actual ET of 52.4 cm yr^{-1}. Hence a small change in either P or ET could cause a large proportional change in N. To summarize, a change in N may be the result of not only a climatic shift, but also a vegetation shift.

In this paper N and γ are steady-state parameters and can be used to discern climatic change only over a certain range of time scales. As mentioned earlier, a steady-state groundwater model is adequate provided that the groundwater system responds faster than the climate can change. In fact, for paleohydrologic applications the steady-state modeling approach is limited more directly by how fast the lake sediments can record a climatic excursion rather than by how fast climate changes. Moreover, for any one lake both the response time and magnitude of lake-level change are functions of the type of climatic or hydrologic shift encountered and the unique hydraulic characteristics of the surface and substrate of the catchment. To conclude, the steady-state results of the groundwater model can be aligned temporally only to those changes in a forcing parameter (climatic or otherwise) that have a periodicity greater than the response time of the sediment records of all the lakes included in the study.

Climatic excursions (as small as individual weather events) that have a periodicity shorter than the response time of the system are those that are combined into the average called the "steady-state" value. The climatic events comprised by a steady-state value are usually not unique, *i.e.*, many different combinations of climatic events may produce the same steady-state value. Hence the interpretation of a steady-state value, or a change in that value, can be problematic. For example, an increase in N could be caused by an increase in the average annual P; alternatively a seasonal redistribution of P that intensifies the spring snowmelt period could also cause an increase in N even if the average annual P remained the same. Because of the precession of the earth's axis during the Holocene, the effects of seasonality are of particular importance. While they will be not be ultimately resolvable using steady-state modeling alone, an iterative model of the soil-moisture budget may be useful in determining the possible seasonal configurations that could result in a given steady-state value.

Climatic implications for 8.5 and 6 ka

The model results for 8.5 ka suggest that N decreased by 60% and γ increased by 20-30 cm yr^{-1} compared to modern values (Fig. 3.6). It is possible that a greater-than-modern T at that time could increase both ET and E, thus accounting for both the decreased N and increased γ. However in light of the northern pine forest that existed in the region immediately prior to 8.5 ka (Jacobson, unpub. data), it seems more probable that Parkers Prairie was at least as cool as today but significantly drier. That is, a reduction in P likewise could account for both the decreased N and increased γ. It is even possible that the T was much lower than today, provided that the decrease in P was large enough to be the dominant control on N and γ values. Alternatively, the seasonality in the Northern Hemisphere was increased at that time compared to today, meaning that in general summers were warmer and winters colder (Kutzbach and Guetter, 1986). Such seasonal intensification could account for an increased γ by increasing summer E more than decreasing winter E, when the lakes would be ice-covered anyway. Further, this intense seasonality could decrease N by causing more of the annual precipitation to fall during the summer months (as might be expected by warmer land masses causing strengthened continental low pressure systems) instead of during the winter and spring snowmelt period.

Although the evidence is equivocal, it is possible that the the water-table elevation jumped up slightly at 8.5 ka because of increased N caused by a reduction in ET as the vegetation shifted from pine forest to prairie. However, by 6 ka the model suggests that both N and γ had increased above their values at 8.5 ka. This combined change in N and γ cannot be easily explained as a simple shift in either P or T alone. The simplest scenario is one in which N increases because of increased P, while at the same time E increases (presumably because of increased T) enough to dominate the γ value. Alternatively, a seasonal redistribution of moisture that decreased summer P and increased late winter and spring P could increase N, but γ would have to be increased by other means.

Thus one interpretation of the lake-level evidence at Parkers Prairie could be that a dry and probably cool climate at 8.5 ka shifted to a warmer and wetter climate at 6 ka. At Elk Lake, 115 km north of Parkers Prairie, Forester *et al.* (1987) postulate a similar climatic sequence: cool and very dry from 7.8 to 6.7 ka, followed by a warmer and wetter period (although still drier than present). In northern Michigan Miller and Futyma (1987) likewise corroborate a dry early Holocene followed by a wetter phase that caused a rise in the water table after 8 ka and initiated organic deposition in many small basins. Paleohydrologic evidence from fluvial systems also seems to support the general hypothesis of a dry early Holocene to about 8 ka, followed by a wetter mid-Holocene (Knox, 1983). [I should remind the reader of the possible 500-yr correction to my radiocarbon ages: 8.5 ka may actually be closer to 8 ka, and 6 ka closer to 5.5 ka.]

The above climatic interpretation is somewhat at variance with the regional pollen evidence, which indicates that the early Holocene (9 ka) was cooler but also wetter than the mid-Holocene (6 ka) in the region of west-central Minnesota (Bartlein *et al.*, 1984). However, the study by Bartlein *et al.* covers the entire upper Midwest and cannot be expected to have the spatial resolution to accomodate each new data point, especially since the Parkers Prairie site lies near the edge of their data set. Winkler *et al.* (1986) found that Lake Mendota in southeast Wisconsin was lower than present from about 6.5 to 3.5 ka. They concluded that reduced precipitation could account for the lower lake level; however, they did not explicitly discount increased evaporation as a factor.

Future directions

Improving the resolution and interpretation of the lake-level modeling technique presented in this paper will require more work on each aspect of the modeling sequence: input to the model, the model itself, and output from the model. Improvement of the input would mean collecting more lake-level data via core-transect methods; new sites with different hydrologic sensitivities should be selected. In the Parkers Prairie sandplain, knowledge of the areal extent of the peatlands for a given past time could remove much of the ambiguity in the regional sensitivity to γ. Of course, the refinement of techniques for discerning lake levels from sediment cores could lead to a vastly improved paleo-lake-level resolution over both time and space.

Adding more geologic realism would improve the groundwater model, and in fact many techniques for doing so already exist. However, the detail of a model is scale dependent, and in general trade-offs exist between the regional scope and local detail of a model (although the analytic-element method nearly entirely circumvents the problems of mixing scales in the same model). Better understanding of the limitations of a Dupuit-Forchheimer (two-dimensional) model would allow more judicious modeling of areas that do not meet the model assumptions as well as a sandplain does; perhaps in thicker and more anisotropic aquifers a three-dimensional model would have to be used. Especially important should be the use of non-steady-state models to investigate the characteristic response times of lake levels and surficial aquifers to different types of climatic changes. Certainly more detailed modeling of fluvial processes could be done, and the interplay between the groundwater and fluvial systems could add important constraints on each other yielding a more unique determination of hydrologic parameters.

Improvement of the model output means to better understand the relation between climatic and modeling parameters and could take two different paths. First, model and climate parameters could be statistically related. That is, N and γ could be related to each other and a host of climatic parameters (P, E, ET, T, *etc.*) via multiple regression or multivariate analyses, provided enough baseline data exists in the literature. Thus with modern data a likely climate could be chosen that might produce the N and γ pair indicated by the groundwater model in its attempt to match paleo-lake levels. Second, the model and climate parameters could be mechanistically related; *i.e.*, the groundwater model could be extended to include a surficial hydrology model and its relation to climatic parameters of interest. In a sandplain, the surficial model would merely partition P into ET and N, probably based on a soil-moisture budget. On other landforms, the surficial model would have to include overland flow. Other refinements might include a lake-evaporation model that allows for changes in the heat storage of the lake, which could be significant for large lakes (much larger than those in the Parkers Prairie sandplain).

SUMMARY AND CONCLUSIONS

The use of an analytic-element groundwater model of the Parkers Prairie sandplain in west-central Minnesota demonstrated the possibility that at 8.5 ka reduced recharge (about 40% of the modern value) was the primary cause of lower lake levels. By 6 ka the model suggested that recharge had increased (to 50 to 80% of the modern value) allowing most lakes to rise in level, but that E - P (lake pumping γ) may have also increased keeping at least one lake nearly dry.

Conclusions

(1) Groundwater can play an important role in determining the pattern of lake-level change across a region. As predicted by Chapters 1 and 2 and then borne out in this chapter, closed-basin lakes far from rivers tend to have larger lake-level changes than do lakes close to rivers.

(2) Vegetation can have a significant effect on regional hydrology and hence also on lake levels. Even small changes in evapotranspiration caused by shifts in the physiognomy of the vegetation could have a disproportionately large effect on recharge to the water table. The genesis and growth processes of peatlands could cause large changes in the area of surfaces exposed to evaporation, and the interaction of peat growth and streamflow may be an important factor in shifting water table elevations.

(3) Lakes respond, to some degree, individualistically to different climatic or hydrologic parameters. For example, the lakes in the Parkers Prairie sandplain had different sensitivities to recharge, lake pumping, and river elevation. Lakes lying in other types of landforms additionally probably have differential sensitivities to overland flow characteristics.

(4) Even though the groundwater model gives quantitative estimates for recharge and lake pumping, at present these values can be related only qualitatively to climatic parameters. It seems clear that the next step should be to mechanistically link the groundwater model to climate (and vegetation, to some degree) via the use of a surficial hydrology model. There are at least two benefits of such a linkage. First, the added complexity should allow a better evaluation of past climate. Each lake is a sensor that is responding uniquely to different climatic signals. These complex responses of lakes should be treated as an asset rather than a liability; expansion of the model to include surface hydrology should allow some of this lake-level complexity to be exploited in

deciphering the details of past climate. Second, the inclusion of surface hydrology would help allow a direct interface with the results of general circulation models (GCMs). This interface is important not only for checking the GCM results for present and past climates, but also for using GCM results for predicting the hydrologic consequences of possible future climates influenced by the greenhouse effect, nuclear or volcanic catastrophe, or ozone depletion.

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