SENSITIVITY ANALYSIS OF A MULTILAYER, FINITE-DIFFERENCE MODEL OF THE SOUTHEASTERN COASTAL PLAIN REGIONAL AQUIFER SYSTEM: MISSISSIPPI, ALABAMA, GEORGIA, AND SOUTH CAROLINA

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U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 87-4108

Atlanta, Georgia

1987

DEPARTMENT OF THE INTERIOR
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U.S. GEOLOGICAL SURVEY
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CONTENTS

ILLUSTRATIONS

TABLES

CONVERSION FACTORS

For use of readers who prefer to use metric (International System) units, conversion factors for inch-pound units used in this report are listed below:

Multiply inch-pound unit	By	To obtain metric unit
foot (ft)	0.3048	meter (m)
foot squared per second (ft^2/s)	0.929	meter squared per second (m^2/s)
foot squared per day (ft^2/d)	0.929	meter squared per day (m^2/d)
inch (in)	25.40	millimeter (mm)
inch per year (in/yr)	25.40	millimeter per year (mm/yr)
mile (mi)	1.609	kilometer (km)

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Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called "Mean Sea Level of 1929."

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ABSTRACT

A sensitivity analysis was made on a multilayer finite-difference regional flow model developed for the Southeastern Coastal Plain aquifer system. It was made on both the steady- and transient-state model input parameters. The results can be used to assess the degree of confidence in the calibrated values of these parameters.

The sensitivity of the model was tested by changing the calibrated values for five parameters in the steady-state model and one in the transient-state model. The parameters changed under the steady-state condition were those that had been routinely adjusted during the calibration process as part of the effort to match predevelopment potentiometric surfaces and elements of the water budget. The tested steady-state parameters include: recharge, riverbed conductance, transmissivity, confining unit leakance, and boundary location. In the transient-state model, the storage coefficient was adjusted. The sensitivity of the model to changes in the calibrated values of the above parameters was evaluated with respect to the simulated response of net base flow to the rivers and the mean value of the absolute head residual. To provide a standard measurement of sensitivity from one parameter to another, the standard deviation of the absolute head residual was calculated.

The steady-state model was shown to be most sensitive to changes in rates of recharge. When the recharge rate was held constant, the model is more sensitive to variations in transmissivity, especially updip in the interstream divide areas where hydraulic gradients are the steepest. Near the rivers, the riverbed conductance becomes the dominant parameter in controlling the heads. In this area, the model is more sensitive to changes in riverbed conductance than it is to comparable changes in transmissivity and confining unit leakance. Change in confining unit leakance has little effect on simulated base flow, but greatly affects head residuals, especially where confining units are thin or their vertical hydraulic conductivity is large. As shown by tests performed on the A3 model layer, the model is relatively insensitive to changes in the location of no-flow boundaries and to moderate changes in the altitude of constant head boundaries.

The storage coefficient was adjusted under transient conditions to illustrate the model's sensitivity to a change in storativity. The model is less sensitive to an increase in storage coefficient than it is to a decrease in storage coefficient. As the storage coefficient decreased, the aquifer drawdown increases, and as a result of a relative flattening of the gradients towards the rivers, the base flow decreased. The opposite response occurred when the storage coefficient was increased.

The results of the sensitivity analysis suggest that the values of the calibrated model parameters are for the most part centrally clustered within the range of reasonable values that might be used to represent the physical system. For the updip and middip areas of the system, the calibrated parameters provide simulated heads that are within about 30 feet of published head data. In the downdip areas of the system, where little data are available for calibration, simulated potentiometric surfaces probably match actual water levels within 50 feet. The simulated responses to changes in the calibrated values suggest that the model's ability to simulate actual conditions deteriorates as departures from the calibrated values increase.

INTRODUCTION

The Southeastern Coastal Plain aquifer system is being studied as part of the U.S. Geological Survey's Regional Aquifer-System Analysis (RASA) program, a series of investigations that present a systematic, unified regional overview and assessment of the hydrogeologic and geochemical conditions of the Nation's major aquifer systems. The area of this investigation is in the southeastern United States and includes part of the Coastal Plain of South Carolina, Georgia, Alabama, and Mississippi (fig. 1). The Southeastern Coastal Plain aquifer system consists of siliclastic rocks of Cretaceous to Holocene age, that crop out in adjacent bands, except where they are overlapped by younger strata. A major objective of this study is to examine the pattern of ground-water flow within the network of regional aquifers whose physical boundaries extend beyond political subdivisions and to simulate this flow by the use of a digital computer.

A sensitivity analysis is an essential part of the model calibration procedure, which makes it possible to evaluate the confidence associated with the calibration of input parameters. If variations in a given input parameter produce only a minor change in the predicted response, the model is relatively insensitive to changes in that parameter; therefore, relatively little confidence should be placed in the calibrated value of that parameter. The sensitivity of the model was evaluated by observing the response of the simulated water levels and base flows to changes in the calibrated status of the model inputs. The simulated water-level response was recorded statistically as the arithmetic mean of the absolute head difference and as the standard deviation of this head difference relative to those simulated under calibrated conditions.

Purpose and Scope

The purpose of this report was to describe the sensitivity of a multilayer, regional ground-water flow model to changes in its input parameters and boundary conditions. A graphical approach was chosen for this report, wherein simulated head and base-flow distributions are plotted as dependent variables against the input parameter that provide them. The graphical approach is justified because of the limited data available for calibration standards and the inherent error associated with regionalized data sets. Accordingly, head profiles were probably the most descriptive way to illustrate a change in the regional flow system. The statistical items discussed in this report are limited to the mean and standard deviation of the absolute head residual that results from using noncalibrated input data.

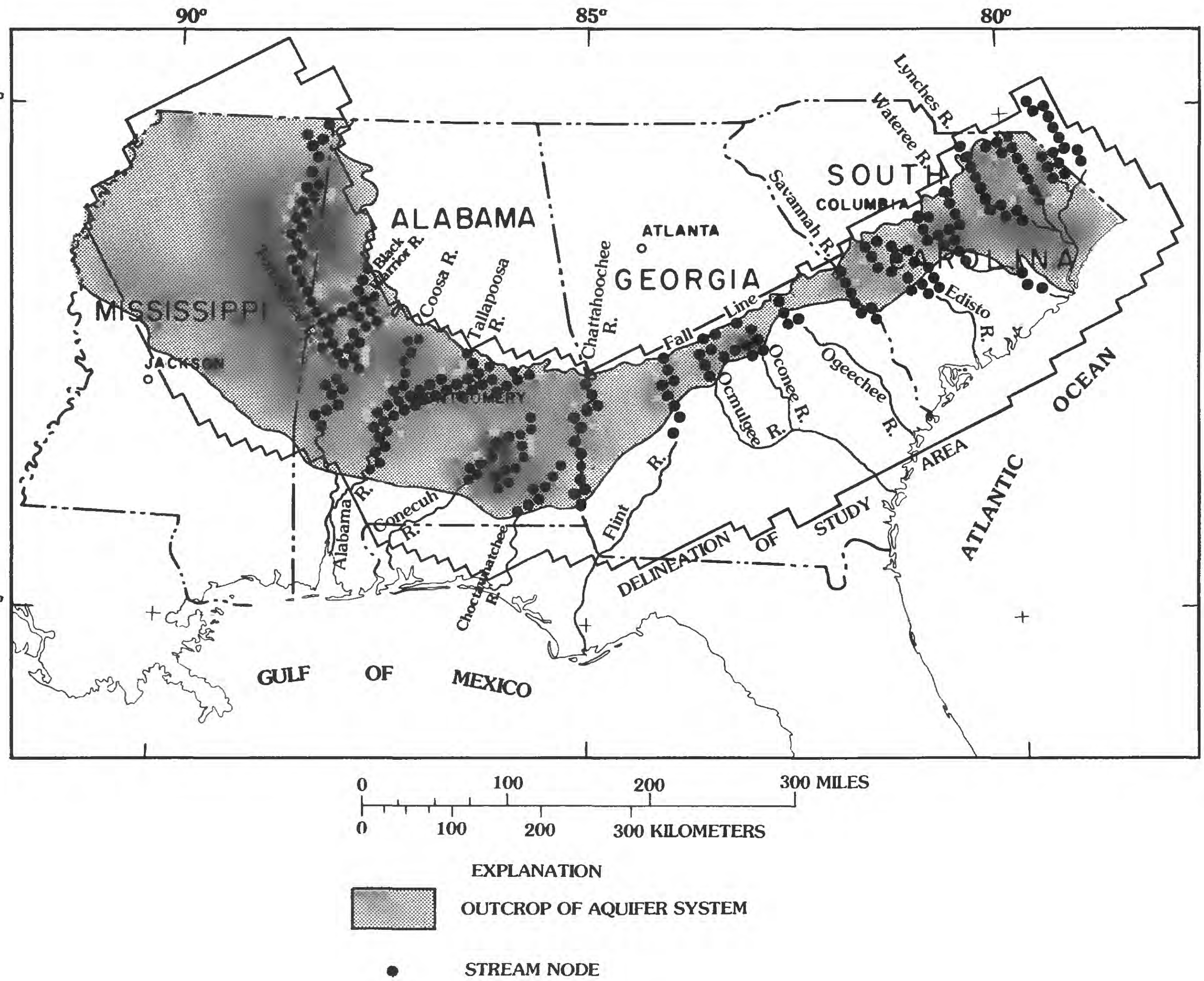

Figure 1.—Location of study area, major streams, and outcrop area of the Southeastern Coastal Plain aquifer system, and stream nodes in the regional model.

In the steady-state analysis, head profiles were drawn longitudinal and transverse to the regional flow direction to give an areal perspective of the model's response to a change in recharge, riverbed conductance, transmissivity, and confining unit leakance. These profiles were selectively chosen in certain areas of Mississippi, Alabama, Georgia, and South Carolina that were known to respond greatly to the adjusted parameters (fig. 2). In the transient-state model analysis, hydrographs of major pumping centers that have widespread drawdown were compared to the simulated response hydrographs when subjected to a change in storage coefficient.

Geohydrologic Setting

The Southeastern Coastal Plain aquifer system whose hydrogeologic framework was mapped by Renken (1984; written commun., 1987) is comprised of a wedge-shaped sequence of unconsolidated to poorly consolidated clastic and carbonate rocks that dip gently seaward from the Fall Line, except in Mississippi where they dip southwest and west towards the Mississippi River. These rocks are the product of the cyclical advance and retreat of ancient seas during Cretaceous to Holocene time. The fluctuating depositional conditions which resulted from sea level changes, regional uplift, and subsidence caused profound changes in the lithic character of the rocks that have been deposited; accordingly, the inherent hydraulic character varies greatly from place to place. Coastal Plain rocks are underlain by metamorphic, igneous, and sedimentary rocks of Paleozoic and early Mesozoic age (Wait and Davis, 1986) that are considered to be impervious to ground-water flow.

Aquifers of the Southeastern Coastal Plain aquifer system consist primarily of coarse- to fine-grained sands that in places contain sandstone, gravel, and minor limestone beds. Confining units that separate the regional aquifers consist of clay, mudstone, siltstone, shale, and chalk.

Recharge to the Southeastern Coastal Plain aquifer system originates as precipitation in the outcrop areas. Annual precipitation in the outcrop area averages about 50 in (Barker, 1986, fig. 12). Only a small part of the precipitation that falls on the outcrop penetrates the deep, confined parts of the aquifer system; the majority of the precipitation either discharges to streams in the form of runoff and base flow, or evaporates, or is transpired by plants. Annual averages of runoff, including base flow in the outcrop area, are approximately 15 in (Barker, 1986).

Ground-water flow in the aquifer system is controlled by variations in hydraulic conductivity and the distribution of recharge and discharge. Water enters the system in updip, outcrop areas where movement is predominately downward. Flow in middip areas occurs along relatively long, nearly horizontal flow paths. Flow in downdip areas is predominately upward because of the decreasing conductivity in the horizontal direction. This, coupled with a rapid buildup in dissolved solids near the downdip limit of permeability, forces water upward into overlying, more permeable zones that contain fresher water.

Acknowledgments

The author wishes to thank Gary Davis, U.S. Geological Survey, Doraville, Ga., for his development of excellent software, allowing the rapid construction of an extensive set of illustrations which provided a comprehensive evaluation of system response.

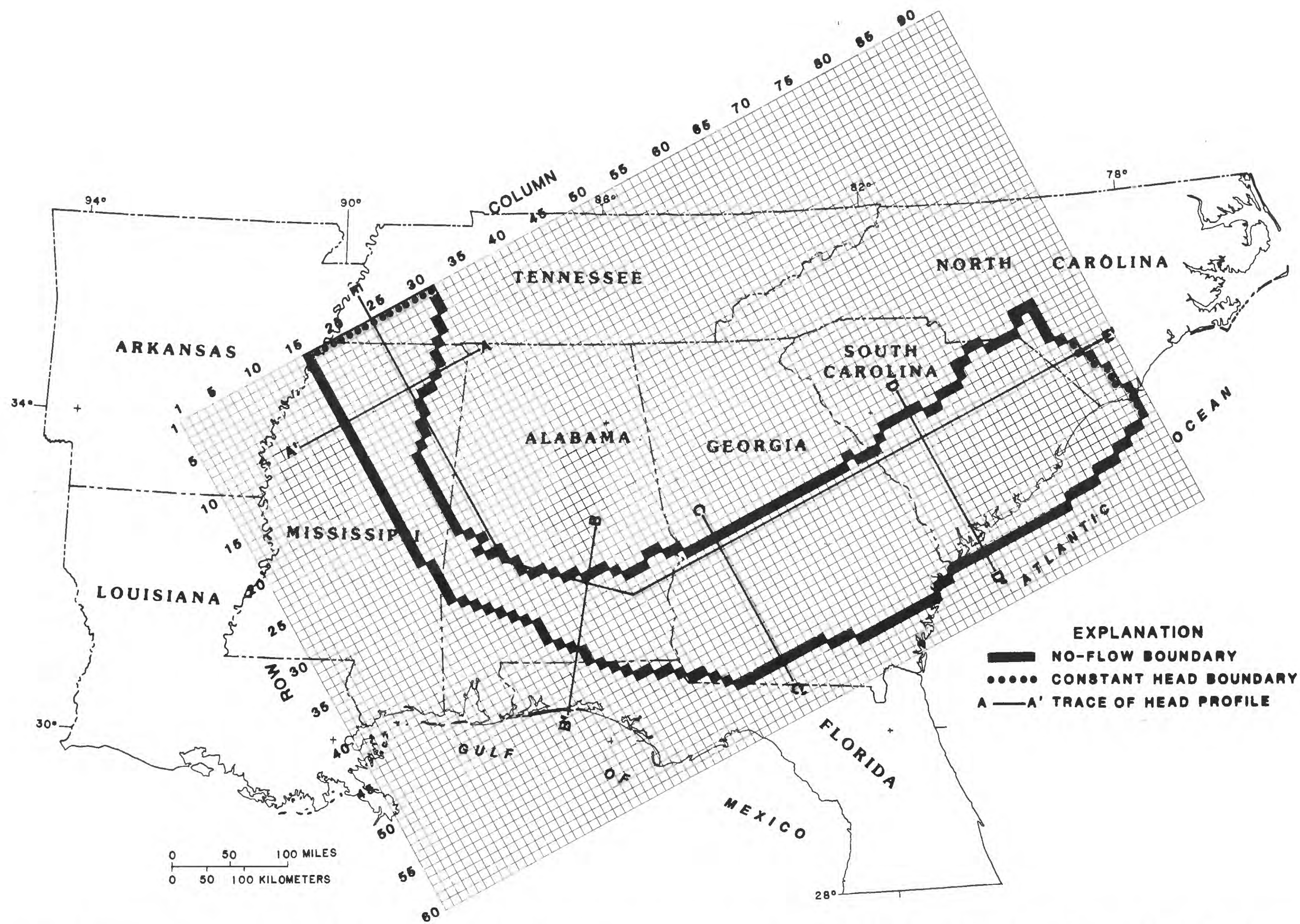

Figure 2.—Finite-difference grid for regional model, showing areal distribution of A3 boundary conditions of the Southeasterr Coastal Plain aquifer system, and the location of longitudinal and transverse head profiles.

MODEL DESIGN

The hydrogeologic framework of the Southeastern Coastal Plain aquifer system was used as a template for the model design. The Southeastern Coastal Plain sediments are divided into seven hydrogeologic units (Renken, 1984); four regional aquifer units separated by three regional confining units that are, in turn, overlain by the Floridan aquifer system in much of the study area (fig. 3).

Input to the regional model is oriented to a finite-difference grid that is a 60-row by 93-column mesh of square blocks, each block being 8 mi to a side. As shown in figure 4, the regional model is comprised of three active aquifers (A2, A3, and A4), overlain by a constant head aquifer (SS), separated by three confining units (C1, C2, and C3). Model layer SS includes the Upper Floridan aquifer (Miller, 1984) and the surficial aquifer.

The regional model used the U.S. Geological Survey's modular three-dimensional finite-difference ground-water flow model (McDonald and Harbaugh, 1984). The strongly implicit procedure (SIP) was used to solve the finite-difference equations of the ground-water flow.

MODEL CALIBRATION

Model calibration is a process through which a numerical representation of the physical system is developed to aid accurate simulation of ground-water flow. The calibrated model can be used to project future conditions in the system in response to anticipated changes in stress. The model was calibrated using a trial-and-error (or indirect) approach because of the limited data available for this study. An indirect approach seeks to improve initial estimates of model input parameters in an iterative way until the simulated heads and base flow match those observed in the field. In this report, the phrases "head" or "hydraulic head" and "water level" refer to the altitude of the water level above sea level.

The steady-state model was calibrated to closely match published predevelopment heads (Stricker, 1984; 1985a and b) and estimated patterns of regional base flow. Model calibration was extended to a transient model which simulates the aquifer system's response to development. The principal input model parameters that were tested include: recharge, riverbed conductance, transmissivity, confining unit leakance, and storage coefficient.

Recharge in the model is provided to nodes that represent the outcrop area of the confined regional aquifer system (fig. 5). The simulated recharge is analogous to water that seeps below the level of small streams that drain the shallow unconfined part of the aquifer system. Regional recharge rates were estimated to average less than 1 in/yr, which is consistent with the difference between the estimated 8 in/yr of total recharge to the entire flow system and the estimated base flow of 7 in/yr to small streams (Wait and others, 1986).

Riverbed conductance is the principal control on stream-aquifer leakance. The regional scale of the model and dimensions of its grid-block of 8 mi to a side limits the simulation of base flow to only the major river drains that are part of the regional ground-water system. Model input values for the riverbed conductance were calibrated through an iterative process to reflect the net

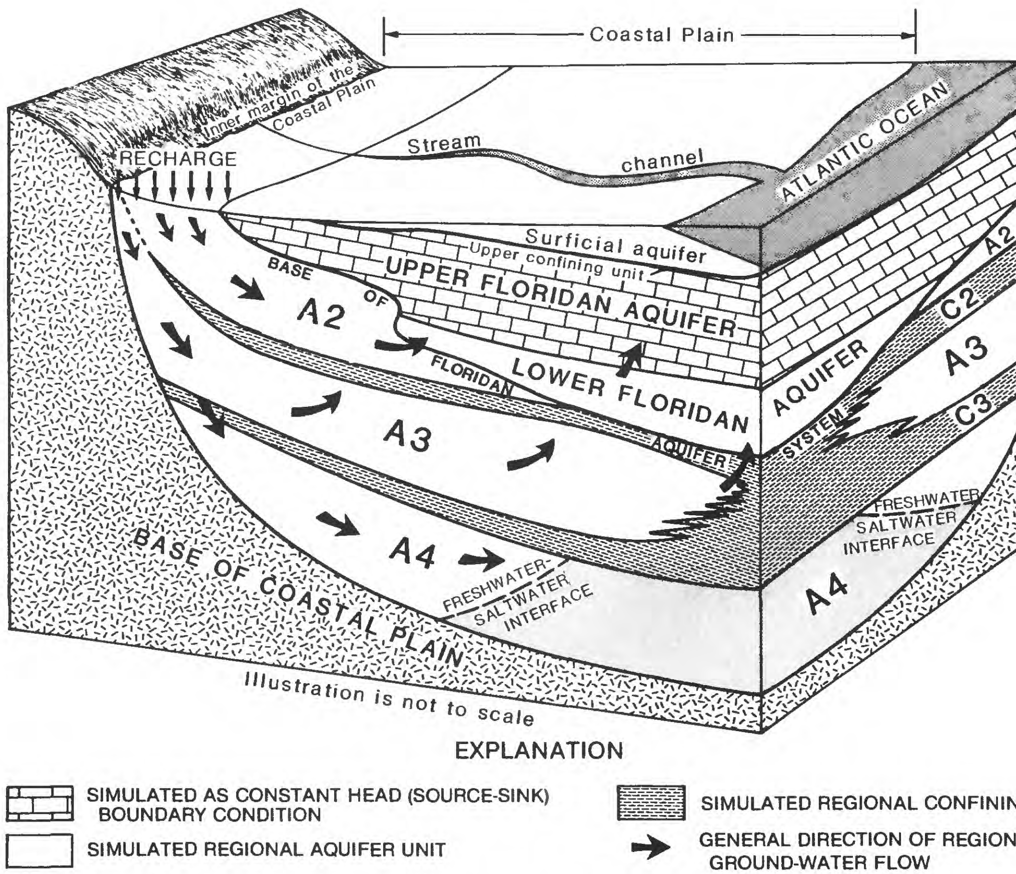

Figure 3.—Relation between hydrogeologic framework and simulated flow direction along a hypothetical dip section through Georgia.

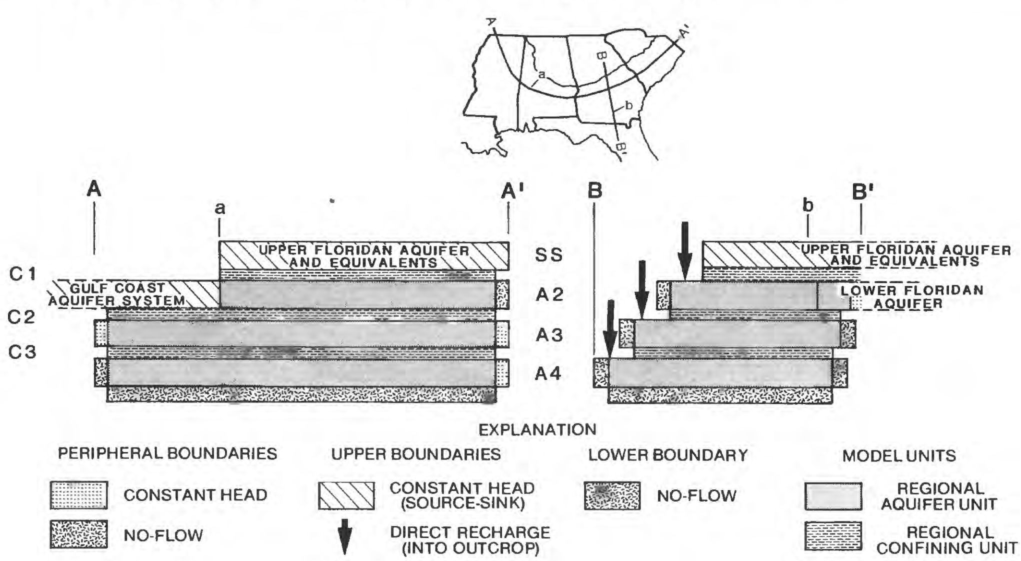

Figure 4.—Schematic strike section (A-A') and dip section (B-B') through simulated Southeastern Coastal Plain aquifer system, showing translation of hydrogeologic framework into model units and the vertical distribution of simulated boundary conditions.

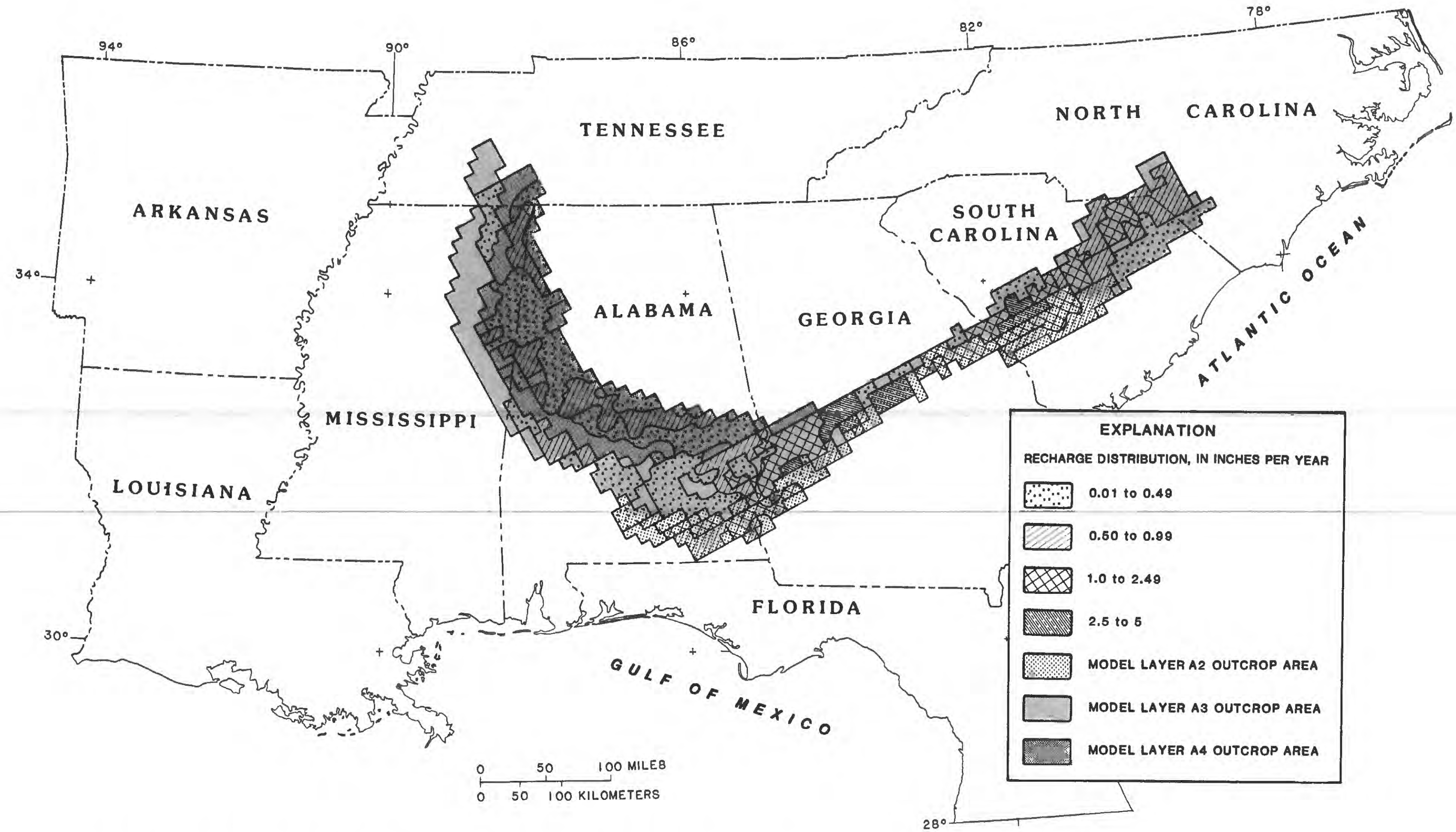

Figure 5.—Areal distribution of calibrated model recharge rates.

effect of streambed geometry and permeability on stream-aquifer leakance. Calibrated riverbed conductance values in the model range from 0.03-0.70 ft^2/s, and average about 0.10 ft^2/s (table 1).

Table 1.--Comparison of average riverbed conductance values for the major drains in the Southeastern Coastal Plain aquifer system.

River	Average conductance in feet squared per second in model layer		
	A2	A3	A4
Tombigbee R.	-	0.046	0.088
Sipsey R.	-	-	.056
Black Warrior R.	-	-	.117
Alabama R.	-	.243	.149
Cahaba R.	-	.065	.078
Coosa R.	-	-	.270
Tallapoosa R.	-	-	.033
Pea R.	0.039	.124	-
Conecuh R.	.118	.268	-
Choctawhatchee R.	.147	.041	-
Chattahoochee R.	.120	.269	.250
Flint R.	.100	.640	-
Ocmulgee R.	.300	.076	-
Oconee R.	.183	.040	-
Ogeechee R.	.040	.200	-
Savannah R.	.533	.330	-
S. Edisto R.	.340	.675	-
N. Edisto R.	.269	.367	-
Wateree R.	.100	.233	-
Lynches R.	-	.475	-
Thompson R.	-	.300	-
Peedee R.	-	.417	-

Estimates of aquifer transmissivity were improved through calibration of the regional model. Calibrated transmissivities range from about 1×10^{-04} to about 5×10^{-01} ft^2/s, and average about 1×10^{-01} ft^2/s (fig. 6).

Confining unit leakance is the ratio of the confining unit vertical hydraulic conductivity and thickness. The original estimates were average values over large areas of the model. These values were calibrated through trial and error adjustment to simulate observed head differentials among model layers. Calibrated leakance values range from about 5×10^{-16} to 1×10^{-08} 1/s, and average about 5×10^{-10} 1/s (fig. 7).

Boundary conditions in the model reflect both physical and hydrological conditions near the limits of the Southeastern Coastal Plain aquifer system. No-flow boundaries are used to simulate the northern boundary of the aquifer system at the Fall Line and the presumed absence of significant freshwater flow across that boundary. No-flow boundaries were also used in layers where the

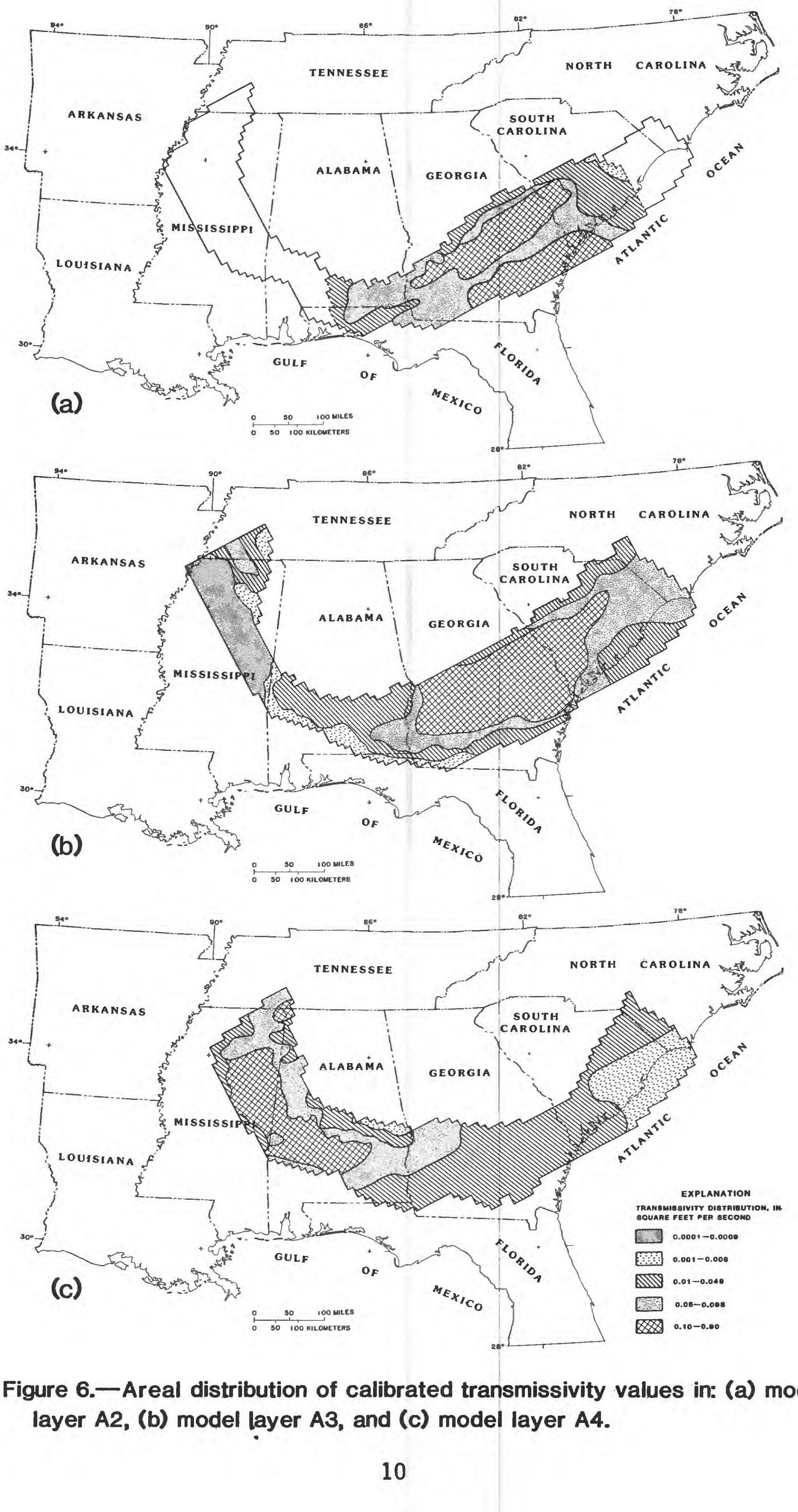

Figure 6.—Areal distribution of calibrated transmissivity values in: (a) model layer A2, (b) model layer A3, and (c) model layer A4.

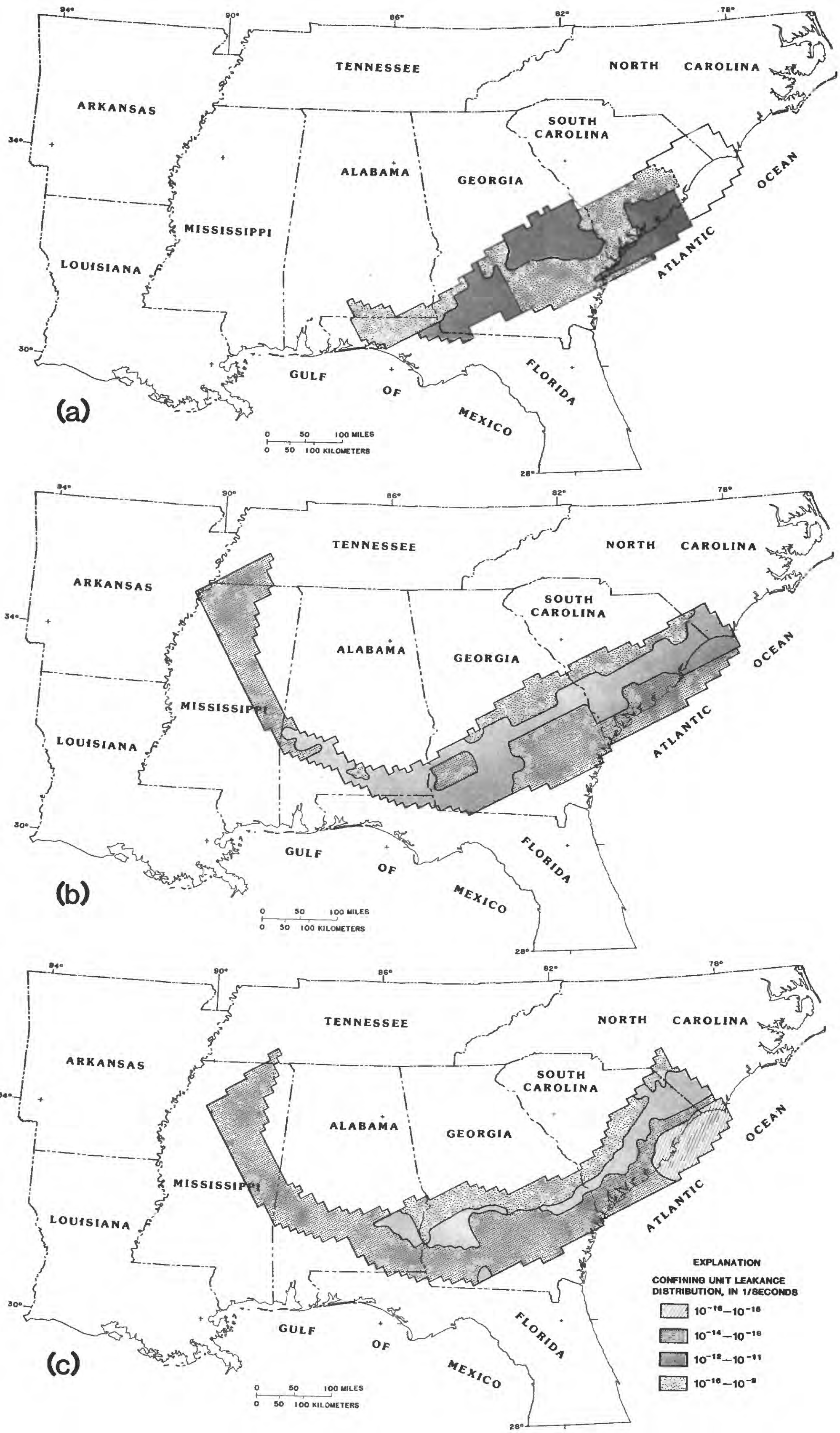

Figure 7.—Areal distribution of calibrated confining unit leakance values in: (a) model layer C1, (b) model layer C2, and (c) model layer C3.

aquifer system was truncated along ground-water divides. In areas where this truncation was not suitable, constant head boundaries were used. Hydraulic heads simulated by adjacent RASA models (Bush, 1982) are incorporated as constant heads atop parts of the Southeastern Coastal Plain model and provide a source-or-sink boundary condition in which water can recharge to the lower layers, or discharge up from the lower layers.

The storage coefficients in the transient-state model were varied regionally using larger values where the aquifer crops out, a midrange in the middip area, and lowest values in the extreme downdip areas (fig. 8). This general gradation is consistent with the expected grain-size distribution in the wedge of clastic sediments on the Coastal Plain (Renken, in press).

METHOD OF STUDY

In the steady-state model, the sensitivity analysis with respect to transmissivity and confining unit leakance was conducted independently in each layer; model sensitivity with respect to change in recharge and riverbed conductance was tested simultaneously in all layers. In the transient-state model, the sensitivity analysis to change in storage coefficient was conducted simultaneously in all layers; only the overall model response, estimated by the average response of an individual layer response, was recorded. The effect of a change in the areal distribution of any of these parameters was not investigated.

The sensitivity of model results to specifications of its boundary conditions was tested for the A3 model layer only and limited to the location of the no-flow boundary nodes and the altitude at the constant head boundaries (fig. 2). Unlike the other model layers, the downdip limits of model layer A3 are based on a qualitative estimate of permeability, which is based on the lithologic character of Coastal Plain rocks rather than the location of a freshwater/saltwater interface (R. A. Renken, U.S. Geological Survey, written commun., 1987).

Model sensitivity was recorded as net base flow to the rivers and as the mean value of the absolute head residual from which the standard deviation was calculated. Net base flow is defined as the algebraic sum of the water discharging from the aquifer to streams and water recharging the aquifer from streams. The mean of the absolute head residual was calculated from the absolute differences between the heads in the calibrated model and the simulated alternative heads, and is expressed as:

$$\overline{h} = \frac{\sum_{i=1}^{n} \left| h_o - h_i \right|}{n}$$

where $\overline{h}$ is the mean value of the absolute head residual; h_o is the simulated head under calibrated conditions; h_i is the simulated alternative head; and n is the number of model nodes with simulated head values. The standard deviation is a measure of the absolute variation about the mean, where one standard deviation incorporates 68.26 percent of the spread. The standard deviation, s, was calculated from:

$$s = \sqrt{\frac{\sum_{i=1}^{n} (h_i - h)^2}{n-1}}$$

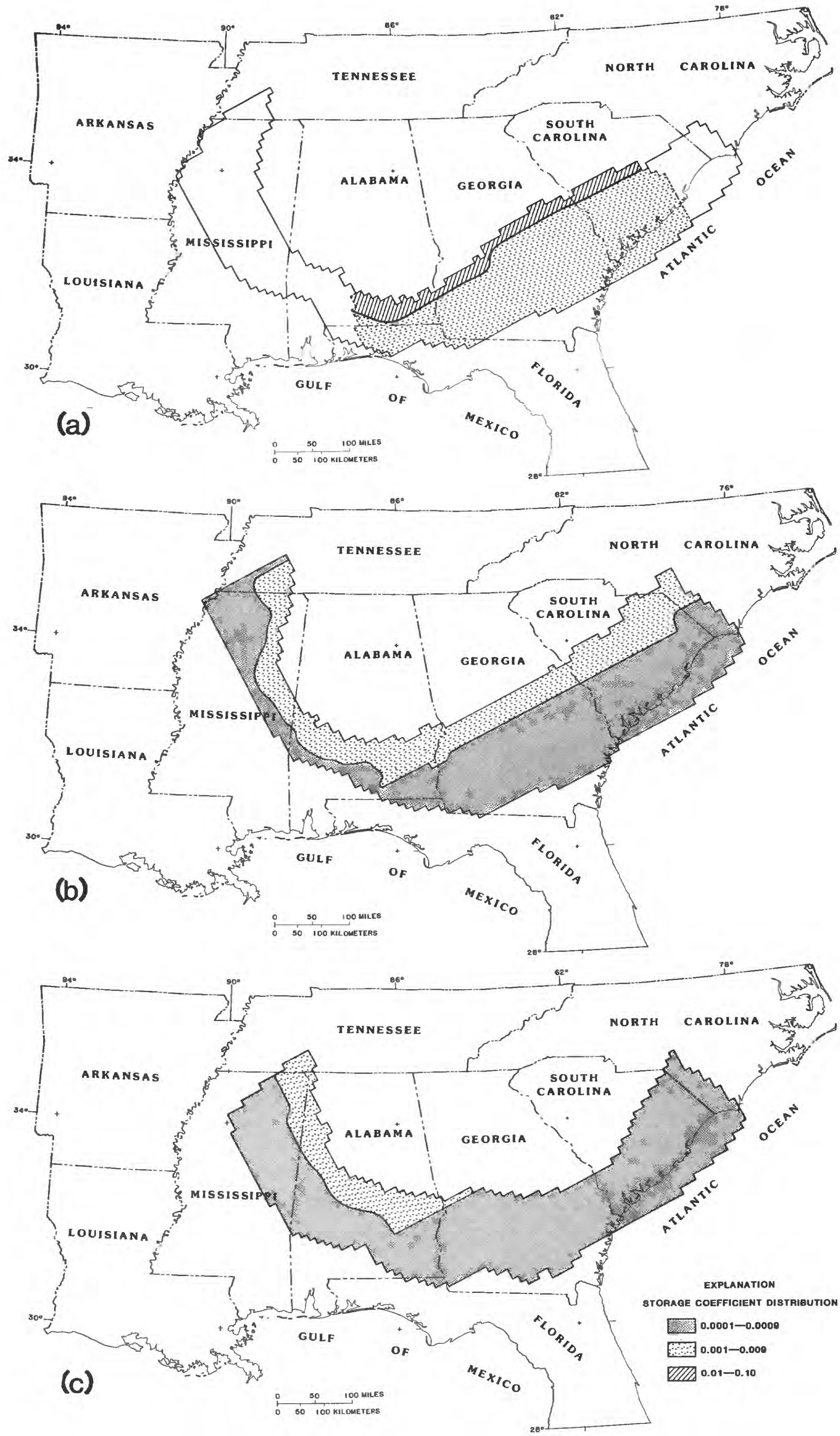

Figure 8.—Areal distribution of calibrated storage coefficients in: (a) model layer A2, (b) model layer A3, and (c) model layer A4.

Each of the input parameters were varied independently by as much as three orders of magnitude greater or less than their respective calibrated values in an attempt to reach the model's threshold response to base flow and head residual. The term "threshold" is defined as the point in the model's response where a further change in the parameter results in little or no change in simulated response.

The aquifer response, expressed either as net base flow or head residual, was plotted against the order of magnitude change in input parameter. The resulting response curve was analyzed based on differences and similarities between the threshold response and conditions simulated using calibrated inputs. Once the threshold value is reached, the response curve flattens, after which the model is considered to be relatively insensitive to further changes in that parameter. As the calibrated value approaches the flat part of the response curve, it becomes increasingly difficult to assess the appropriateness of the calibrated value with respect to its physical counterpart. Confidence in the calibrated value is, accordingly, less as the flat part of the curve is approached.

In addition to plotting the aquifer response to base flow and head residual on an individual model layer basis, the overall model response was recorded. That is, net base flow to the rivers was summed in all layers, and the overall mean head residual was averaged for all layers. The overall mean head residual is generally less than the mean head residual in the model layer which is most sensitive to a given change.

For the steady-state model, head profiles were drawn longitudinal and transverse to the general flow direction. Longitudinal head profiles were chosen in northern Mississippi, central Alabama, western Georgia, and central South Carolina and are labeled as profiles A-A´, B-B´, C-C´, and D-D´, respectively (fig. 2). These profiles illustrate the variation in simulated heads from the Fall Line to the downdip limit of flow. Heads in model layer A2 are constant in Mississippi and a portion of Alabama (profiles A-A´ and B-B´). In addition to the longitudinal profiles, a head profile labeled E-E´ in figure 2 was drawn in each of the active model layers (A2, A3, and A4) transverse to the general flow direction. This head profile provides a cross-sectional view of the head distribution and illustrates how local head gradients in the interstream areas respond to a change in a given input parameter.

The altered head distributions illustrated in the head profiles represent the change in head with respect to distance from the updip outcrop limit of an individual model layer resulting from a half an order of magnitude increase and decrease in a given input parameter from the calibrated value. For leakance and transmissivity, such changes do not always produce a significant deviation from the calibrated profile.

In the transient-state model, only the sum of the net base flow to the rivers and the mean head residuals averaged in all model layers was recorded. Hydrographs were also used to illustrate the model's sensitivity to changes in storage coefficient.

SENSITIVITY OF STEADY-STATE MODEL INPUTS

The sensitivity of the steady-state model was tested by changing the value of the input parameters by as much as three orders of magnitude in an attempt to obtain the model's threshold response to that parameter. The threshold is the point in the response where further change in the input parameter results in little or no change in the model's response. The simulated results were recorded in terms of absolute head residual and base flow; the former was used to calculate the standard deviation between the calibrated and altered head distributions. The results of the sensitivity analysis can be used to evaluate the confidence that should be associated with the calibrated model values that are expected to represent the physical system.

Recharge

Calibrated recharge values were simultaneously changed in all layers by one order of magnitude, and the response was recorded in terms of the absolute head residual and base flow (fig. 9). The standard deviation of the absolute head residual was also used to evaluate the model's sensitivity to changes in recharge.

In general, as recharge increases the net base flow to the rivers and water levels increases. In all the layers, the model is more sensitive to an increase in recharge than it is to a decrease in recharge with the difference in the standard deviation between the two changes being approximately a factor of 10. This implies that the model is less sensitive to a decrease in recharge than it is to an increase in recharge with the limited number of river nodes being a major restriction for the flushing of excess water resulting from additional recharge.

The response of the model to the additional recharge is a function of the riverbed conductance, stream stage, and the transmissivity in the area where the recharge is applied. High transmissivity values in and near the outcrop area induce lateral movement of the additional water downgradient, while the riverbed conductance controls the flow of water into the river. Where the riverbed conductance is low, discharge to the rivers is reduced and the additional recharge remains in the updip areas, resulting in an increase in heads. This situation is exemplified in model layer A4, as the updip transmissivity values near the outcrop are relatively high (values range from 0.05 to 0.60 ft^2/s), but the riverbed conductance of the A4 rivers are low (values average less than 0.20 ft^2/s). As a result, the standard deviation of the head residual resulting from an order of magnitude increase in recharge is about 500 ft.

Model layer A3 shows the greatest base flow response to changes in recharge because more rivers cut the outcrop of the hydrogeologic unit represented by this layer, and because the riverbed conductance of the A3 rivers is relatively higher (values average 0.60 ft^2/s) than those along the rivers cutting the other hydrogeologic units.

The head profiles in the longitudinal and transverse flow direction illustrate this conclusion (figs. 10 and 11). In central Alabama (profile B-B´, fig. 10), an increase in recharge in all layers results in a sharp increase in the heads near the outcrop, while a decrease results in only a slight decrease in the heads from their calibrated positions. This indicates that the model is

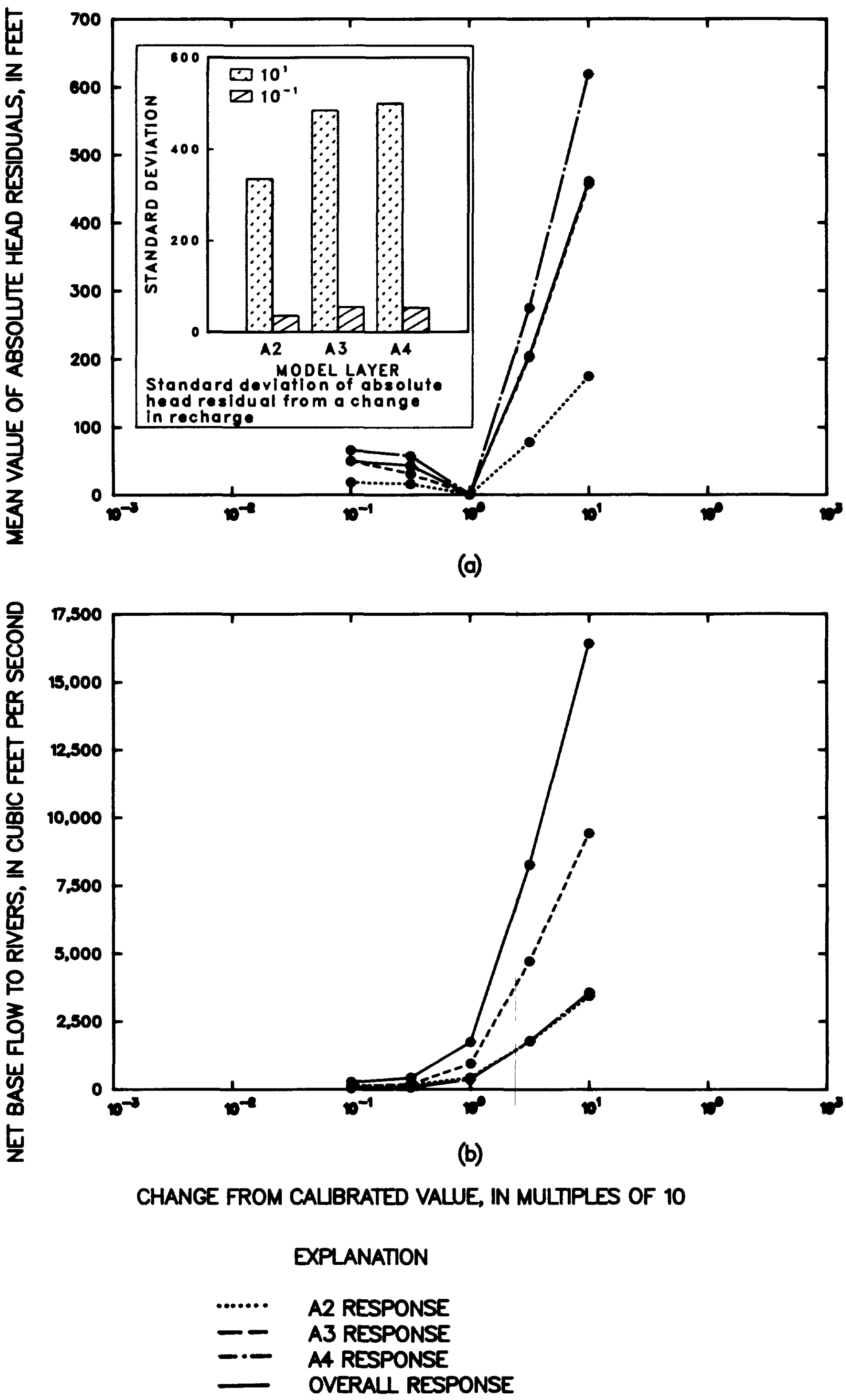

Figure 9.—Model sensitivity to change in recharge with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

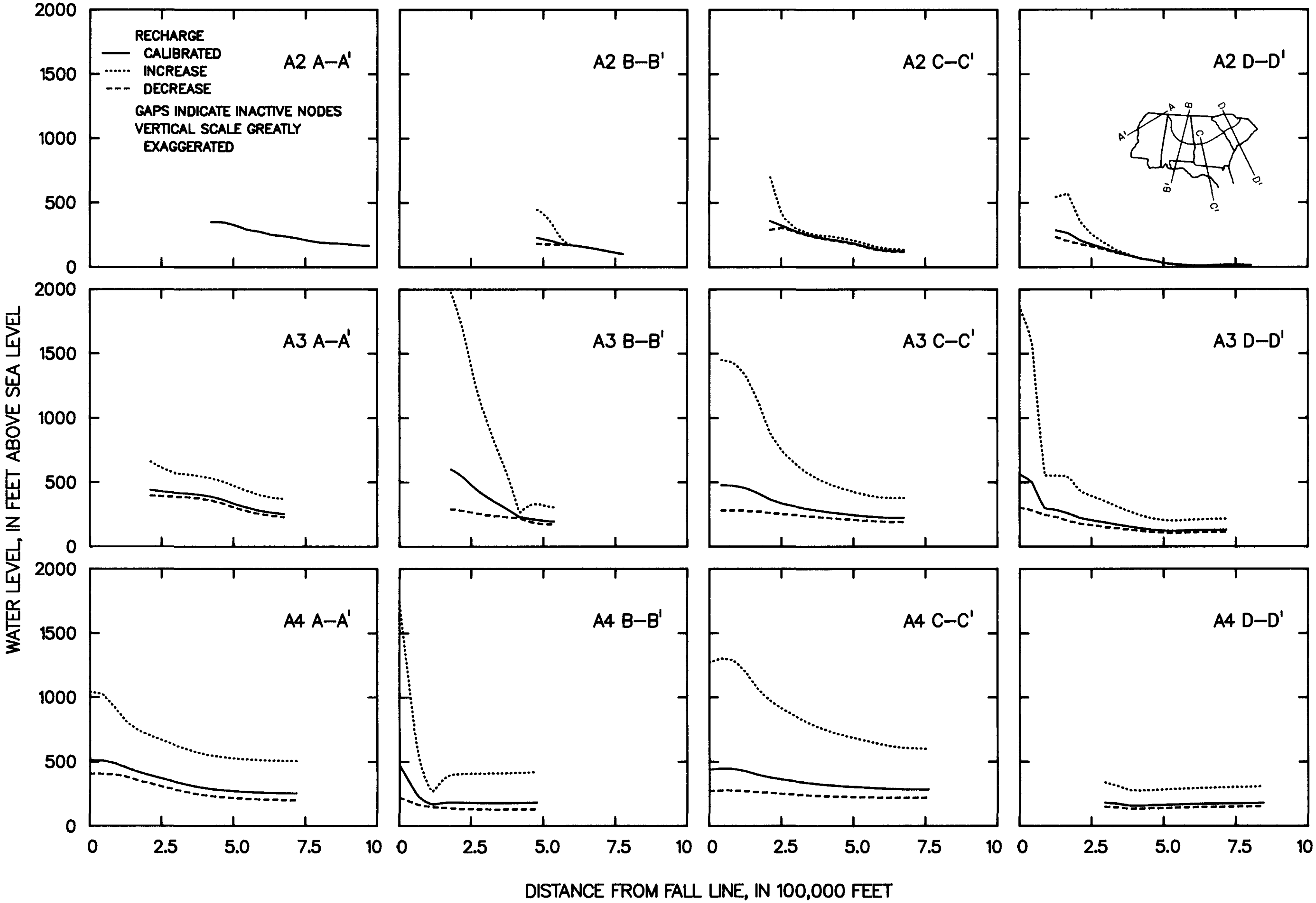

Figure 10.–Longitudinal head profiles from the calibrated recharge and for a half an order increase and decrease in recharge.

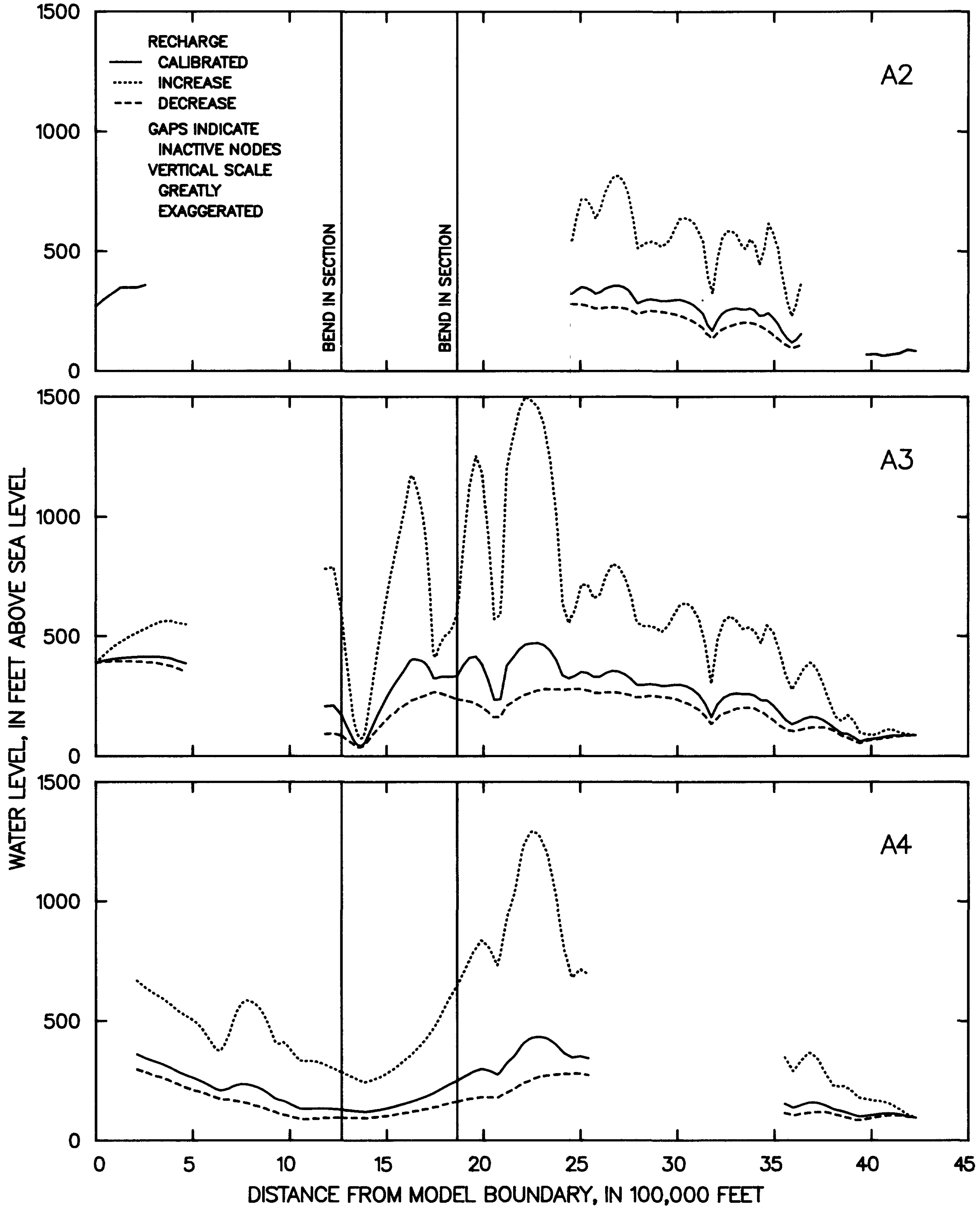

Figure 11.—Transverse head profiles from the calibrated recharge and for a half an order of magnitude increase and decrease in recharge.

able to adjust to a loss in recharge by intercepting the water that would otherwise discharge to rivers. The decrease in base flow (fig. 9b) associated with a decrease in recharge supports this conclusion. The model uses this water to help maintain hydraulic gradients present in the calibrated profile. Despite a significant increase in the aquifer-to-river flow when the recharge is increased, the model must adjust to the additional recharge by raising heads everywhere.

The local head gradients in the interstream area significantly steepen when recharge is increased, but drop to a lesser degree when the recharge is decreased from its calibrated value (fig. 11). This is more apparent in profiles for model layers A3 and A4 than in profiles for model layer A2, as the heads in the latter layer are controlled primarily by the boundary conditions in the overlying source-sink (SS) model layer.

Riverbed Conductance

Calibrated riverbed conductance values were increased and decreased by two orders of magnitude simultaneously for all model layers. The responses with respect to the absolute head residual and the net base flow are shown in figure 12. As the value of the riverbed conductance decreases, less water discharges to the rivers; net base flow decreases, and because water can no longer discharge as freely from the aquifers to the rivers, heads increase.

Head residual response in all model layers is more sensitive to a decrease in riverbed conductance than it is to an increase (fig. 12a), but the opposite is true of base flow (fig. 12b). A decrease in riverbed conductance has the greatest effect on raising heads in model layer A4, while the greatest impact on base flow is seen in model layer A3 due to the overall higher aquifer transmissivity. For example, where the calibrated riverbed conductance is lowered by an order of magnitude, the standard deviation of model layer A4 is 160 ft as compared to the standard deviation of 80 ft for model layer A3.

Figures 13 and 14 show the hydraulic head distribution drawn longitudinal and transverse to the regional flow direction, respectively. All head profiles show a symmetrical deviation from the calibrated profile when riverbed conductance is changed (fig. 13).

The response to changes in the riverbed conductance in model layer A2 is simulated only in the updip sections of the profiles as this is where the rivers are located in the hydrogeologic unit corresponding to this model layer. The downdip heads in model layer A2 are stabilized primarily by the constant heads in the overlying source-sink model layer (SS), and show relatively little response to changes in riverbed conductance.

At profile D-D´ (fig. 13), the calibrated profile for model layer A3 shows a distinct break in slope in the middip range, which is typical of other areas of the model. This situation is perhaps an indication of a transition from relatively high rates of updip recharge (via downward percolation) to relatively small rates of discharge (via diffuse upward leakage), resulting in relatively flat gradients downdip. Because the A3 does not overlie the A4 in extreme downdip parts of this section, the heads in model layer A4 are higher in the downdip areas than they are in the middip areas, as water cannot discharge upward to model layer A3.

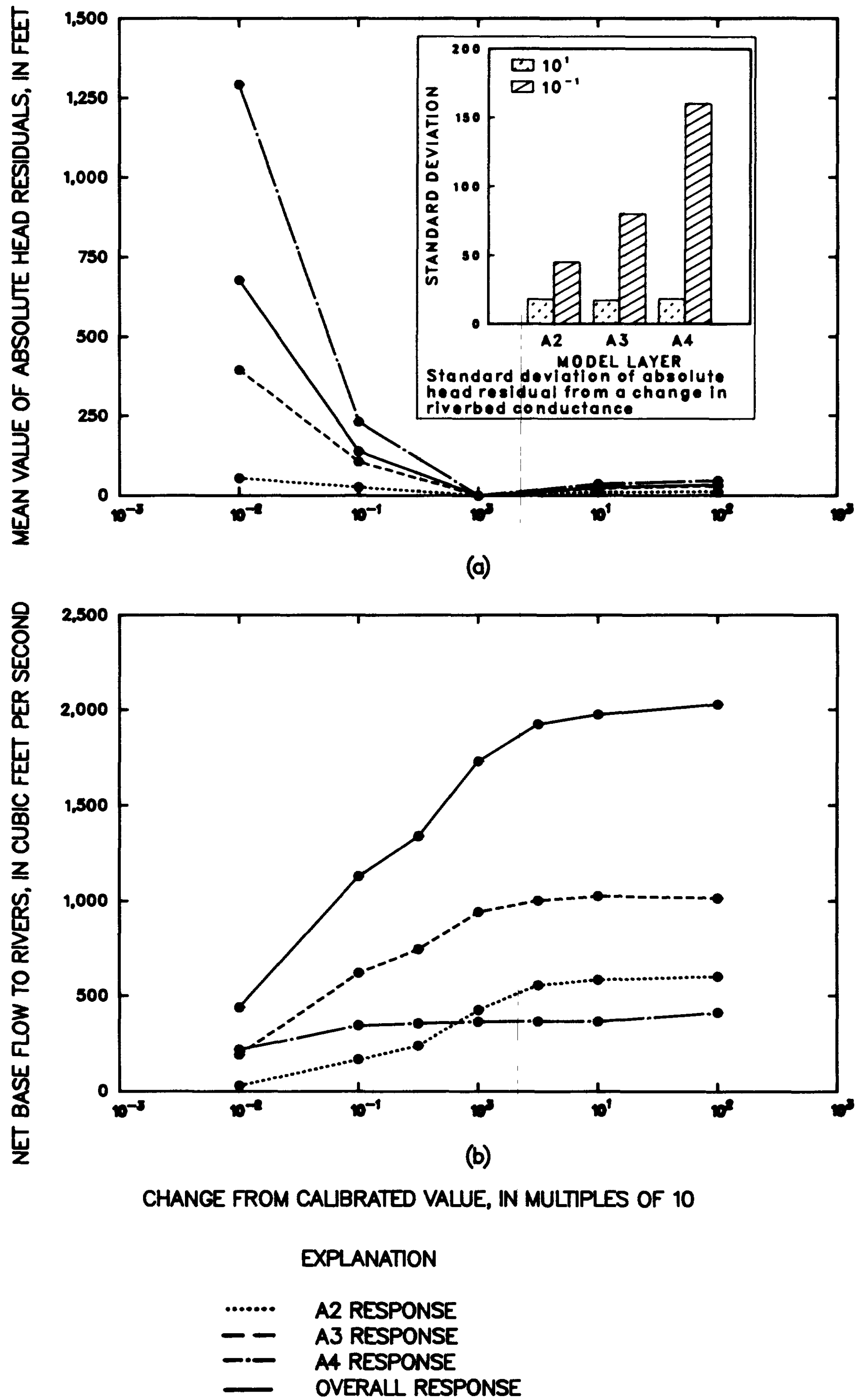

Figure 12.—Model sensitivity to change in riverbed conductance with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

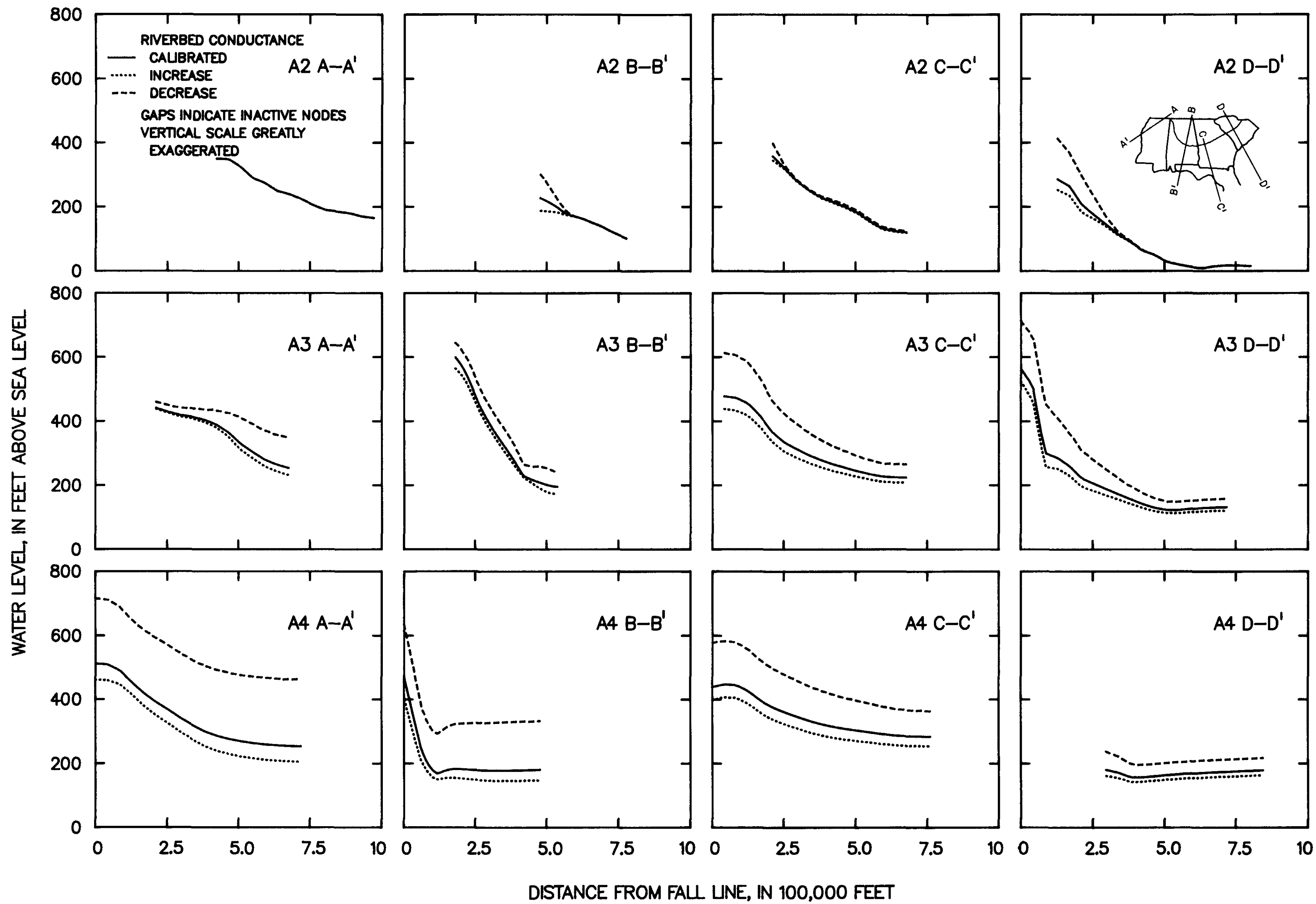

Figure 13.-Longitudinal head profiles from the calibrated riverbed conductance and for a half an order magnitude increase and decrease in riverbed conductance.

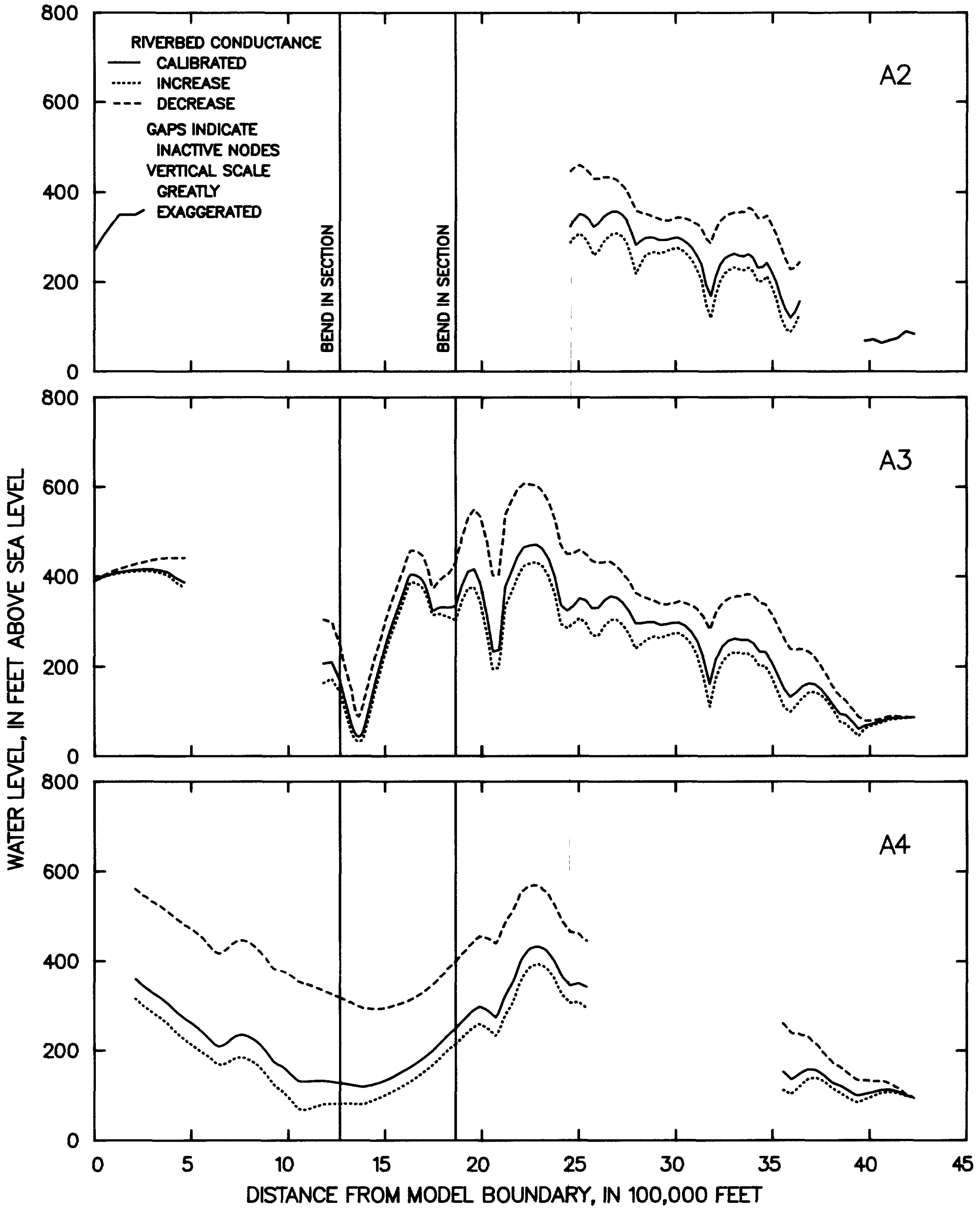

Figure 14.—Transverse head profiles from the calibrated riverbed conductance and for a half an order of magnitude increase and decrease in riverbed conductance.

As the riverbed conductance decreases, flow towards the rivers is decreased, resulting in an extremely high interstream head profile. An increase in riverbed conductance more closely maintains the head gradients that exist under calibrated conditions (fig. 14). In areas where the calibrated riverbed conductances are highest (such as near the Peedee River in fig. 14), an increase in the conductance values has very little effect on reducing the heads. In other areas of the model (such as near the Tombigbee River in fig. 14), where the calibrated conductances are lower, an increase in conductance results in a greater head reduction. This is probably a result of the differences between the transmissivity values adjacent to these rivers which controls the lateral flow of water.

Transmissivity

Calibrated transmissivity values were changed independently in all layers by two orders of magnitude. The standard deviation of the absolute head residual was used to judge the model's sensitivity to a variation in transmissivity.

Head residuals in all model layers are more sensitive to a decrease in transmissivity than to an increase (figs. 15a, 16a, and 17a). The standard deviation of the head residual resulting from a decrease in transmissivity is 3 to 6 times greater than the standard deviation resulting from an increase, with the largest deviation occurring in the layer in which the change was made.

In terms of base flow (figs. 15b, 16b, and 17b), the model is more sensitive in all layers to an increase in transmissivity than it is to a decrease. An increase in transmissivity has the greatest effect on base flow of the rivers in model layer A3, regardless of the layer in which the transmissivity is changed.

In general, the model is more sensitive to a change in transmissivity in the updip areas, where the hydraulic gradients are generally steep, than it is to a change downdip where the gradients are relatively flat. The extent that a change in transmissivity in one layer influences the adjacent layers depends to a large extent on the leakance of the confining units separating the aquifers and the proximity of boundary conditions. As a given confining unit becomes more conductive, either by thinning or by an increase in vertical hydraulic conductivity, the hydraulic connection between adjacent aquifers increases and the effect of a transmissivity change on the aquifers may increase. The proximity to boundary conditions is important in determining the magnitude of response, as shown by model layer A2 (figs. 18 and 19). The head profile in this model layer is controlled primarily by the constant heads in the overlying SS model layer, and as a result, a change in the transmissivity values in model layer A2 has little effect on the heads in this layers or in the underlying layers (fig. 18).

Longitudinal head profiles (figs. 18, 20, and 22) show that as transmissivity increases, water levels updip decrease as more water flows downgradient away from the outcrop which, in turn, causes an increase in heads in the mid and downdip areas. This decrease in updip heads causes less water to discharge into the streams, thus reducing the net base flow to the rivers. As transmissivity decreases, the longitudinal head profile updip increases, as downgradient flow is retarded; base flow decreases due to the restricted lateral movement of water towards the rivers. These effects are more pronounced in aquifer layers A3 and A4 (figs. 20 and 22) than in aquifer layer A2 (fig. 18).

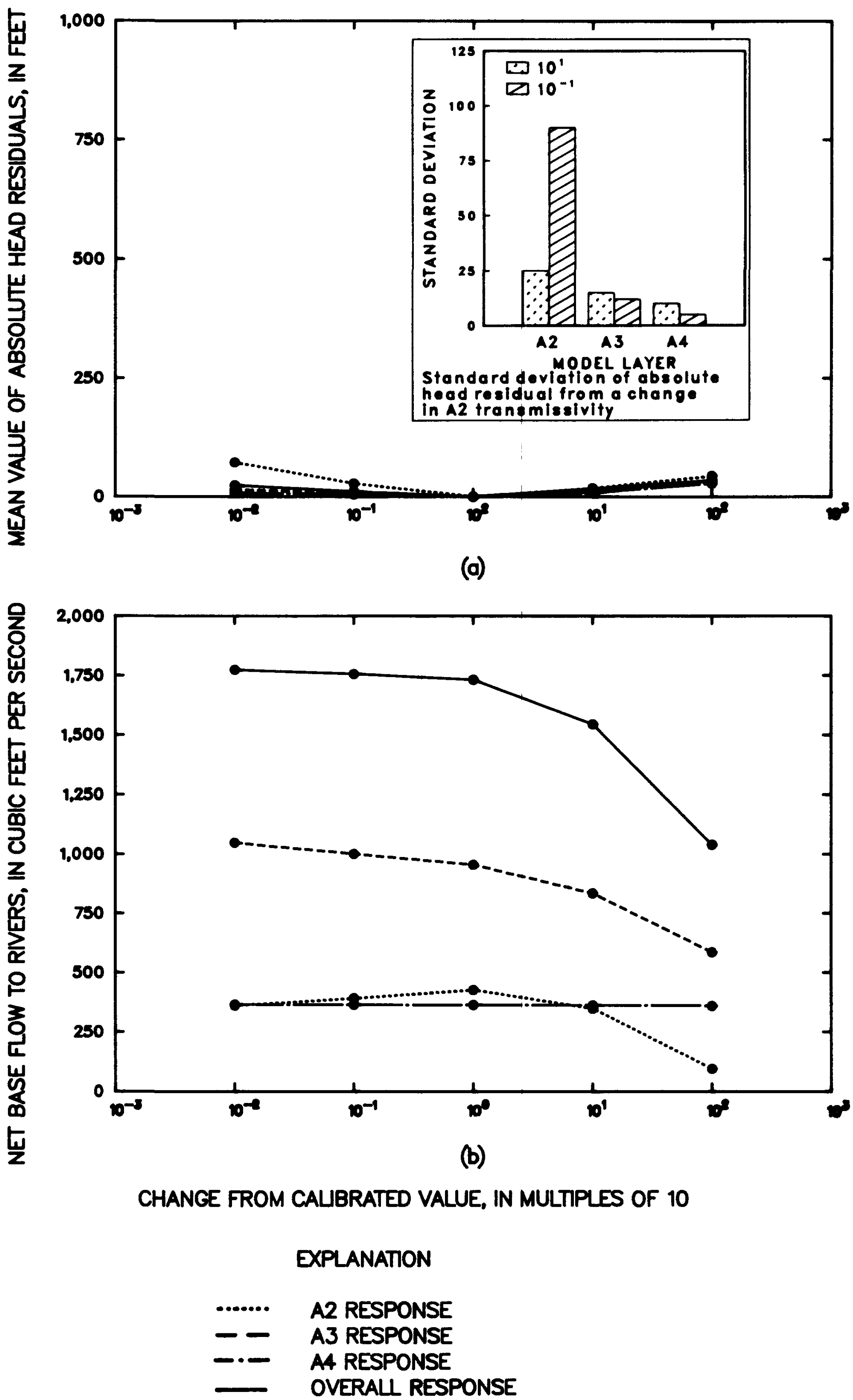

Figure 15.—Model sensitivity to change in A2 transmissivity with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

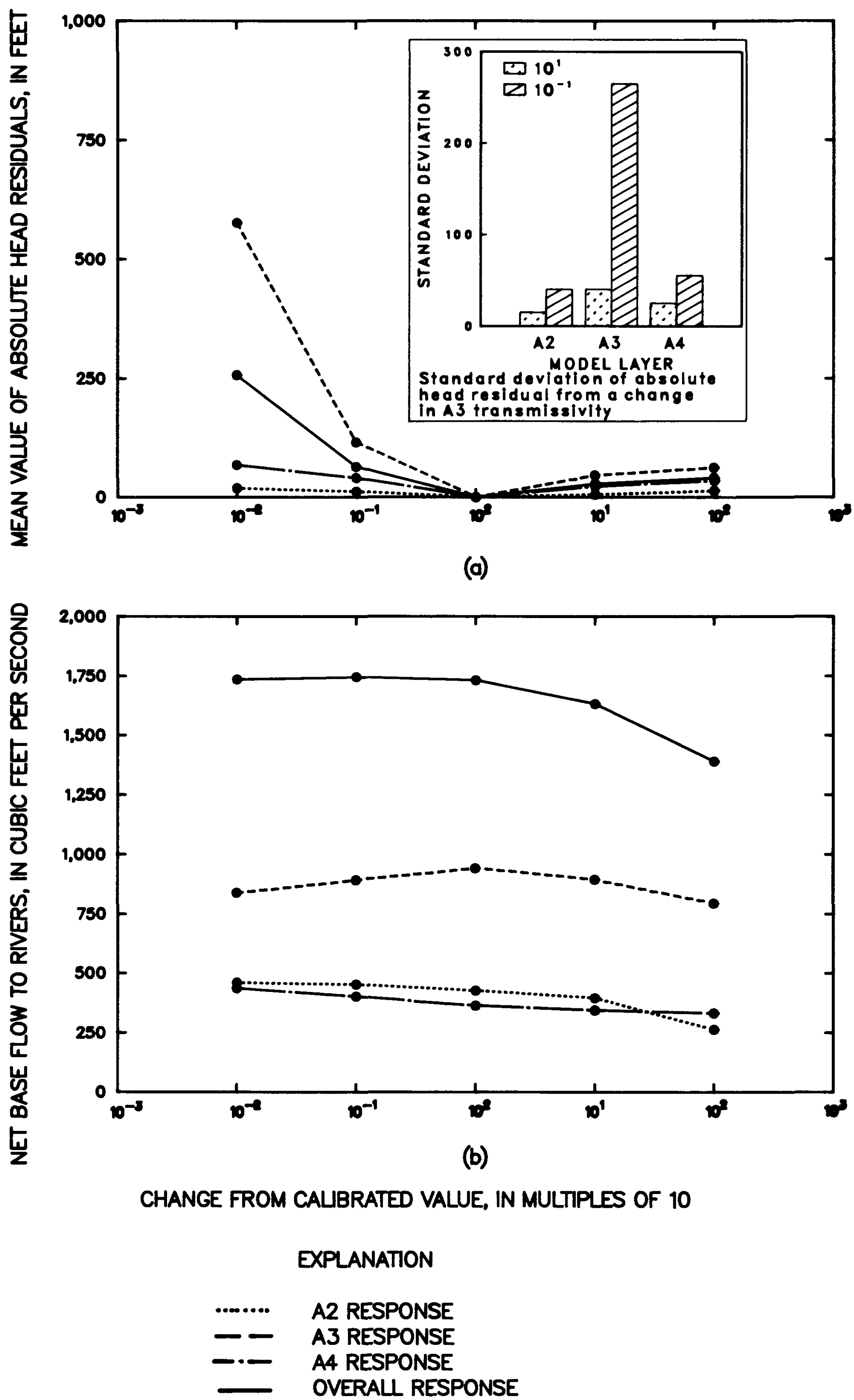

Figure 16.—Model sensitivity to change in A3 transmissivity with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

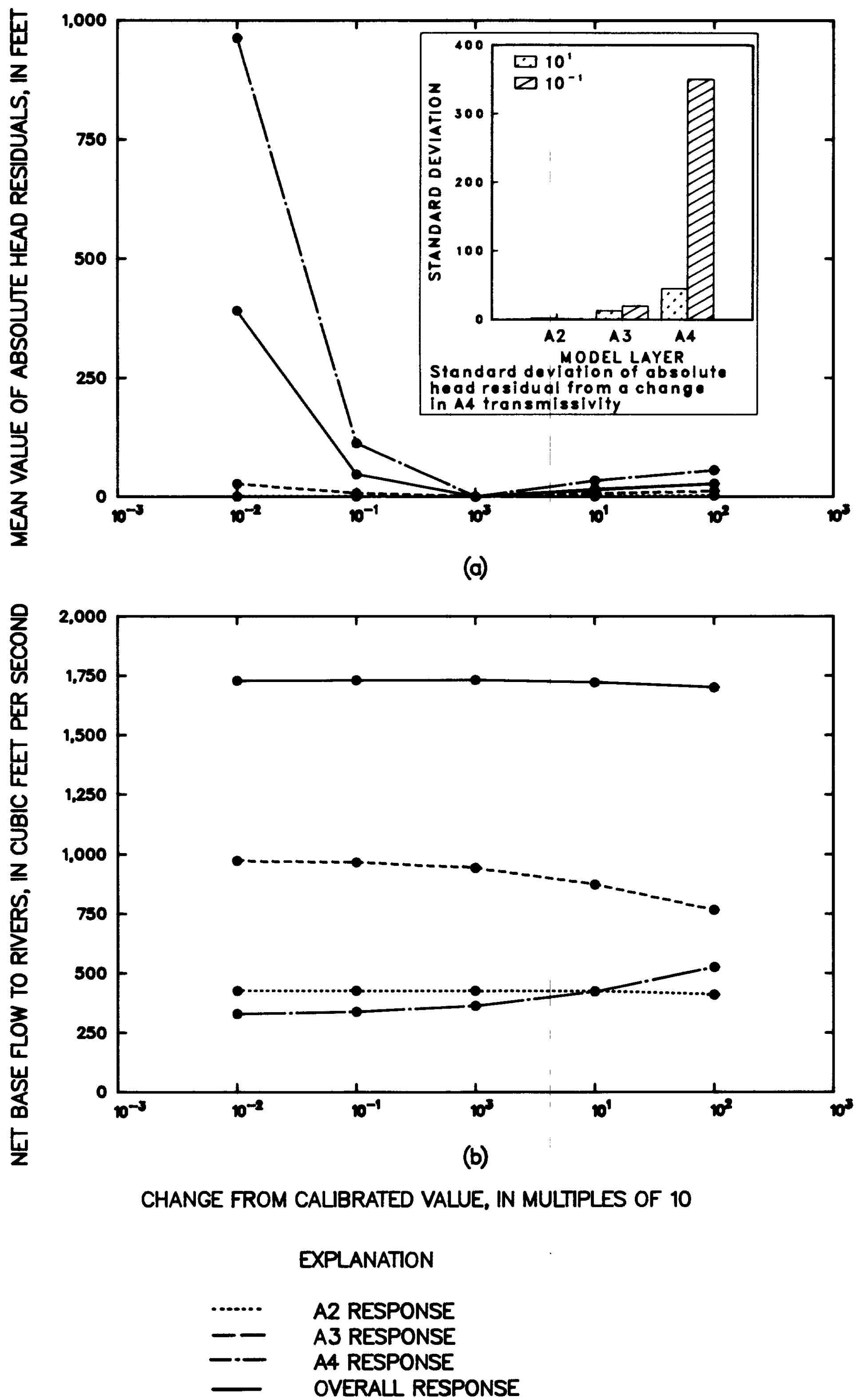

Figure 17.—Model sensitivity to change in A4 transmissivity with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

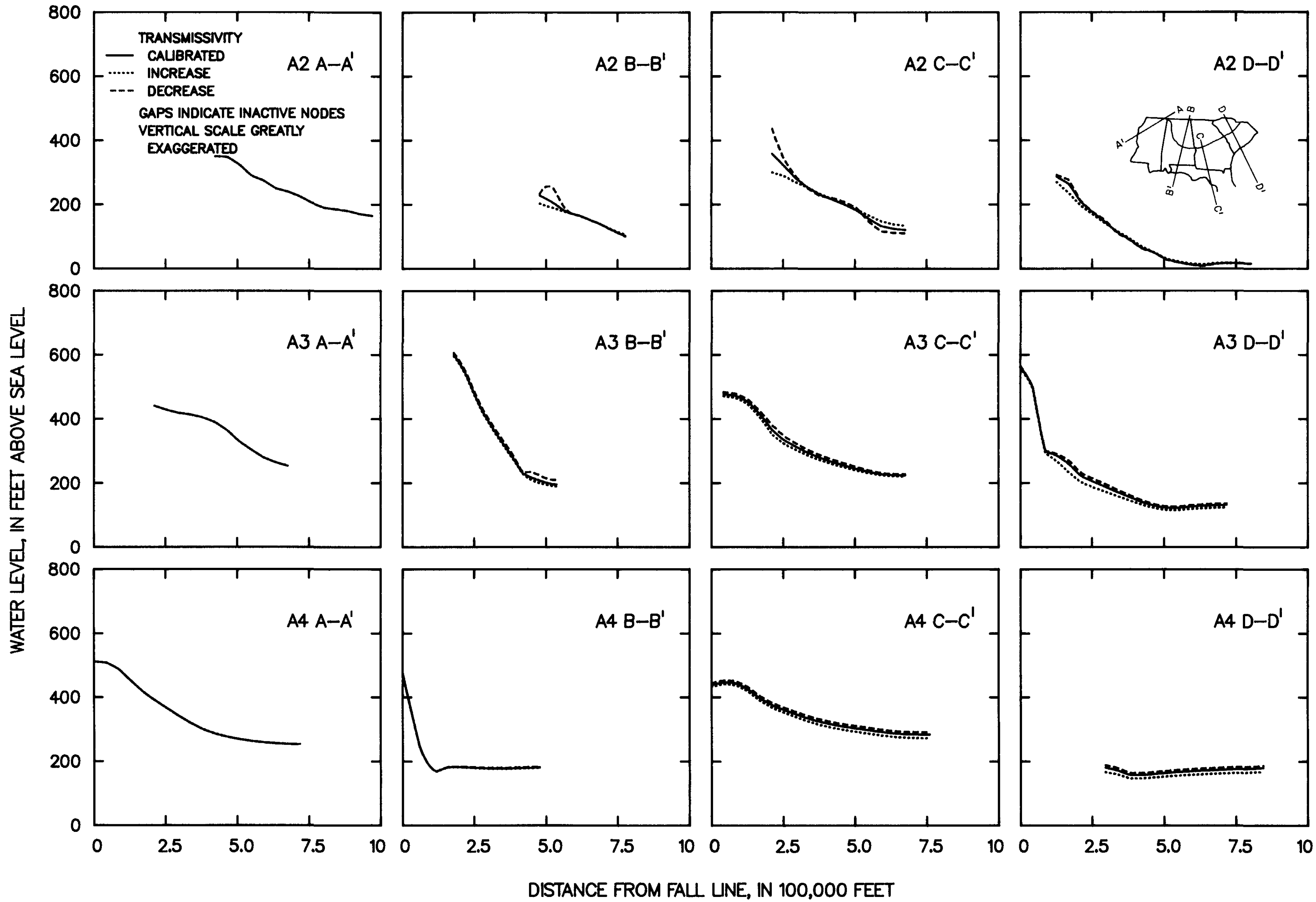

Figure 18.—Longitudinal head profiles from the calibrated A2 transmissivity and for a half an order of magnitude increase and decrease in A2 transmissivity.

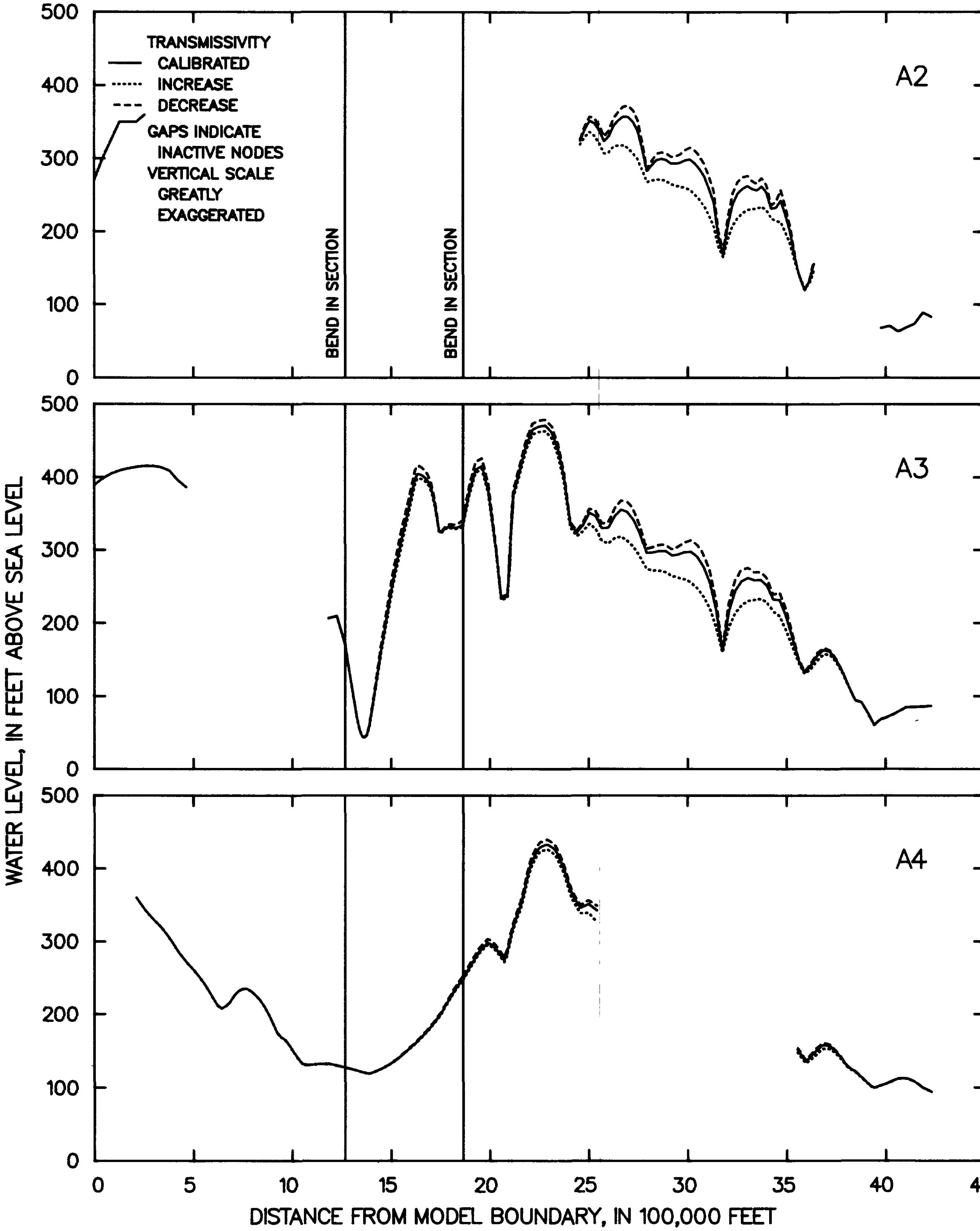

Figure 19.—Transverse head profiles from the calibrated A2 transmissivity and for a half an order of magnitude increase and decrease in A2 transmissivity.

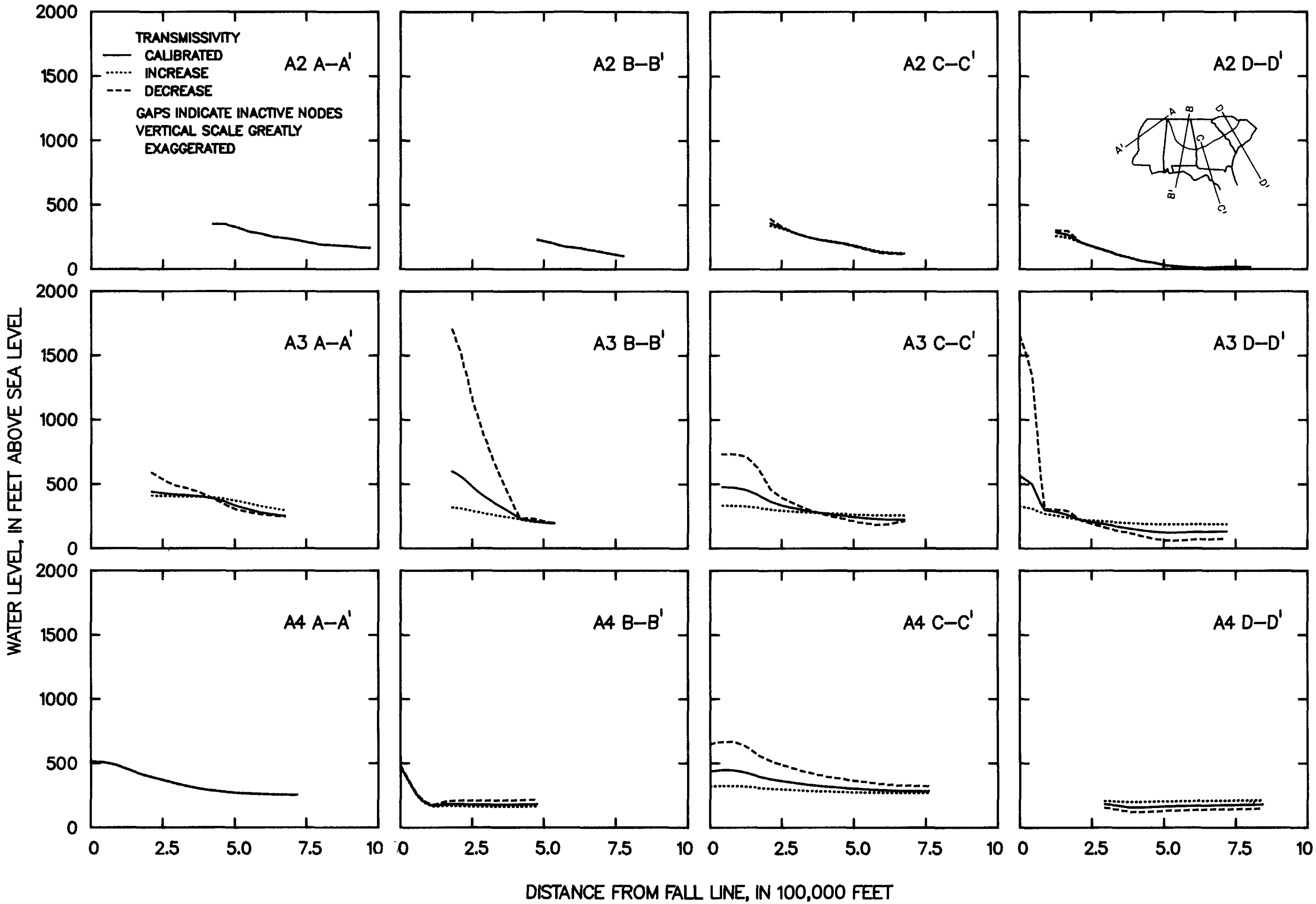

Figure 20.—Longitudinal head profiles from the calibrated A3 transmissivity and for a half an order of magnitude increase and decrease in A3 transmissivity.

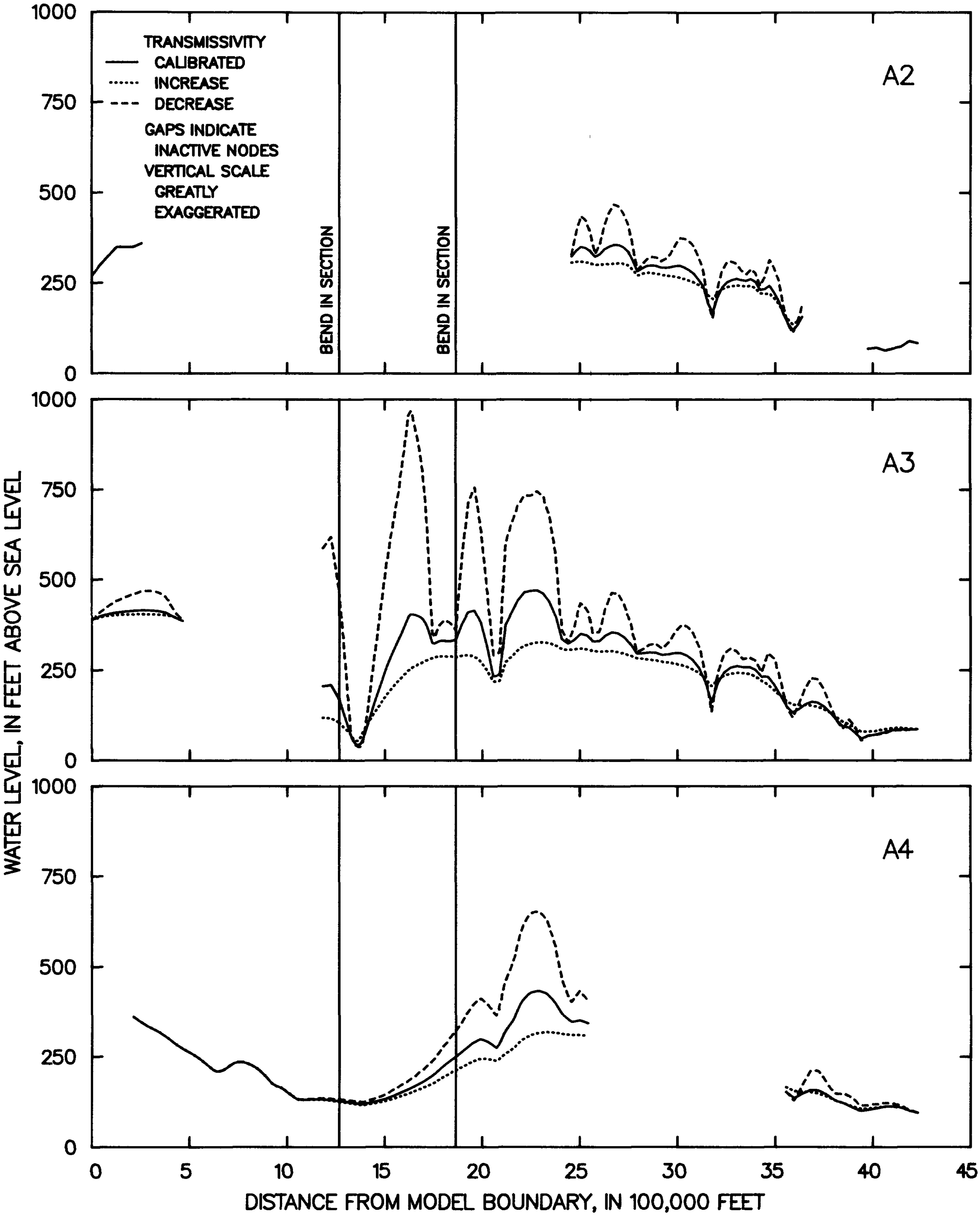

Figure 21.—Transverse head profiles from the calibrated A3 transmissivity and for a half an order of magnitude increase and decrease in A3 transmissivity.

The longitudinal head profiles resulting from changing the transmissivity in model layer A3 a half an order of magnitude are shown in figure 20. A change in the head profile in model layer A3 has the greatest influence on the longitudinal profile of model layer A4 in western Georgia (profile C-C', fig. 20). Here, the confining unit separating model layers A3 and A4 is relatively thin updip and its leakance is as high as 1 x 10^{-10} 1/s. As transmissivities in model layer A3 decrease, there is a corresponding increase in head in the updip areas of this layer; as a result, the vertical flow between model layers A3 and A4 increases as the model simulates a new head equilibrium.

The transverse head profiles resulting from a change in transmissivity in model layer A3 illustrate the influence of topography on the head profiles (fig. 21). The local head gradients are much steeper with a decrease in transmissivity than those with the calibrated transmissivity. However, the increased gradient is not enough to overcome the effects of decreased transmissivity, and the net result is decreased base flow (fig. 16). A change in transmissivity in model layer A3 influences the head profile in model layer A2 between the Flint and Wateree Rivers (fig. 21) because the confining unit separating the two layers is relatively leaky there (1 x 10^{-10} 1/s).

Changes in the transmissivity of model layers A3 and A4 show a similar effect on the heads in the adjacent layers between the Chattahoochee and Flint Rivers (figs. 21 and 23) as this is where the confining unit separating the two is relatively leaky (1 x 10^{-10} 1/s). West of this area, there is very little communication between the aquifers as the confining unit thickens and vertical leakance decreases to as low as 1 x 10^{-14} 1/s.

Confining Unit Leakance

The calibrated confining unit leakance values were changed independently in all layers by three orders of magnitude and the response of the absolute head residual and base flow was recorded (figs. 24, 25, and 26). The standard deviation of the absolute head residual index due to an increase and decrease in leakance are not suitable sensitivity indicators. A change in the leakance of any confining unit by as much as two orders of magnitude results in a standard deviation of less than 40 ft in any one layer and this is considered to be within the calibration objectives.

The model is more sensitive to a decrease in leakance than it is to an increase. A decrease in leakance has the greatest effect on the heads in the aquifer underlying the confining unit whose leakance is decreased (figs. 24a, 25a, and 26a). Simulated base flow in the model is relatively insensitive to changes in leakance (figs. 24b, 25b, and 26b). This is because the rivers analyzed for base flow are located updip in and near the outcrop, where the confining units are either absent or very thin. Accordingly, changes in the flow patterns in middip and downdip areas where the confining units are effective has little effect on the flow to rivers in the updip, outcrop areas.

The influence of a change in leakance on adjacent aquifers depends on the relative calibrated leakance of the confining unit and the calibrated head differential, or gradient, between the adjacent aquifers. When a confining unit is thin or relatively conductive, vertical flow between adjacent aquifers is enhanced, while a thick or less conductive confining unit inhibits vertical flow. The head gradient between adjacent aquifers depends on the location and

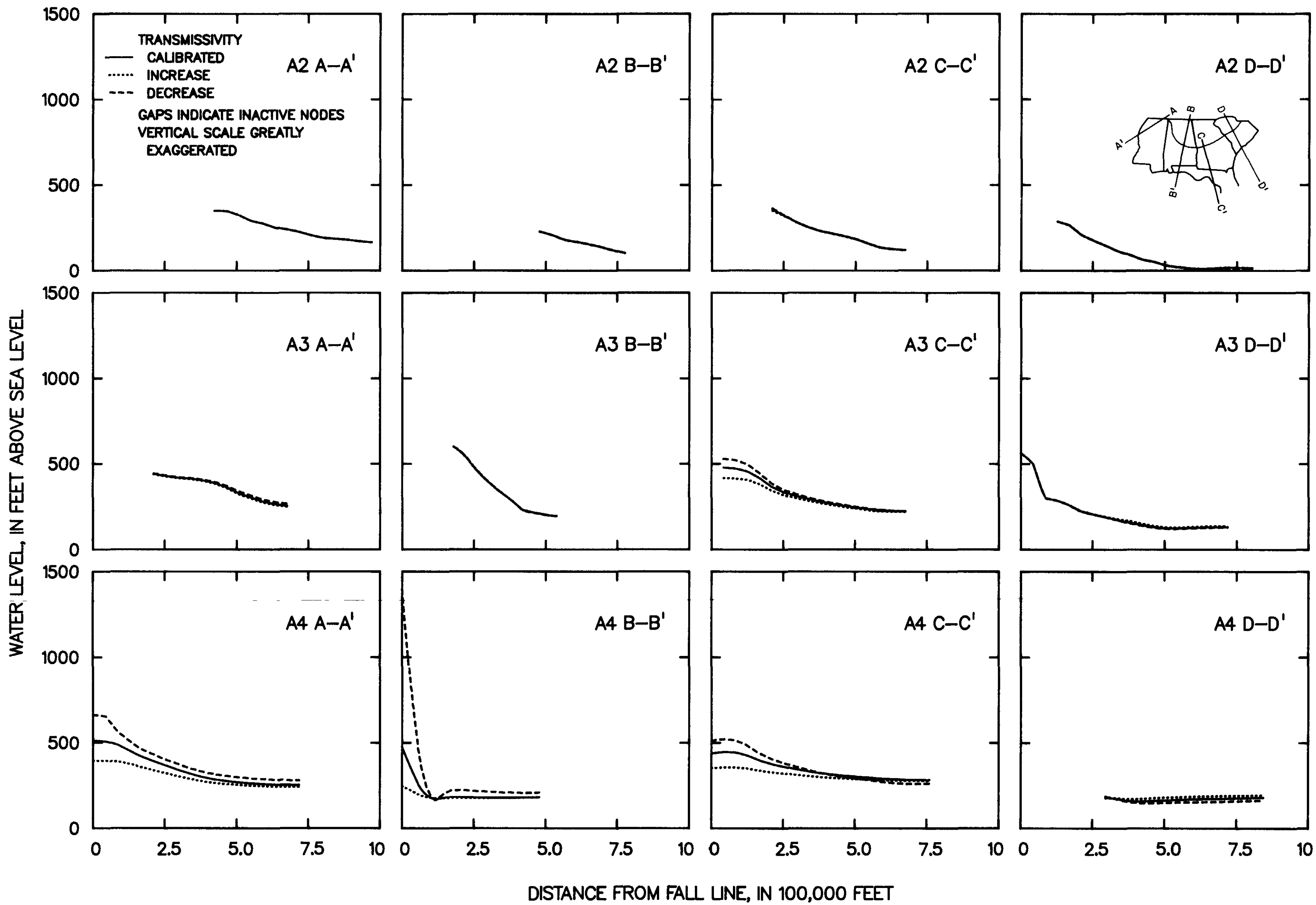

Figure 22.—Longitudinal head profiles from the calibrated A4 transmissivity and for a half an order of magnitude increase and decrease in A4 transmissivity.

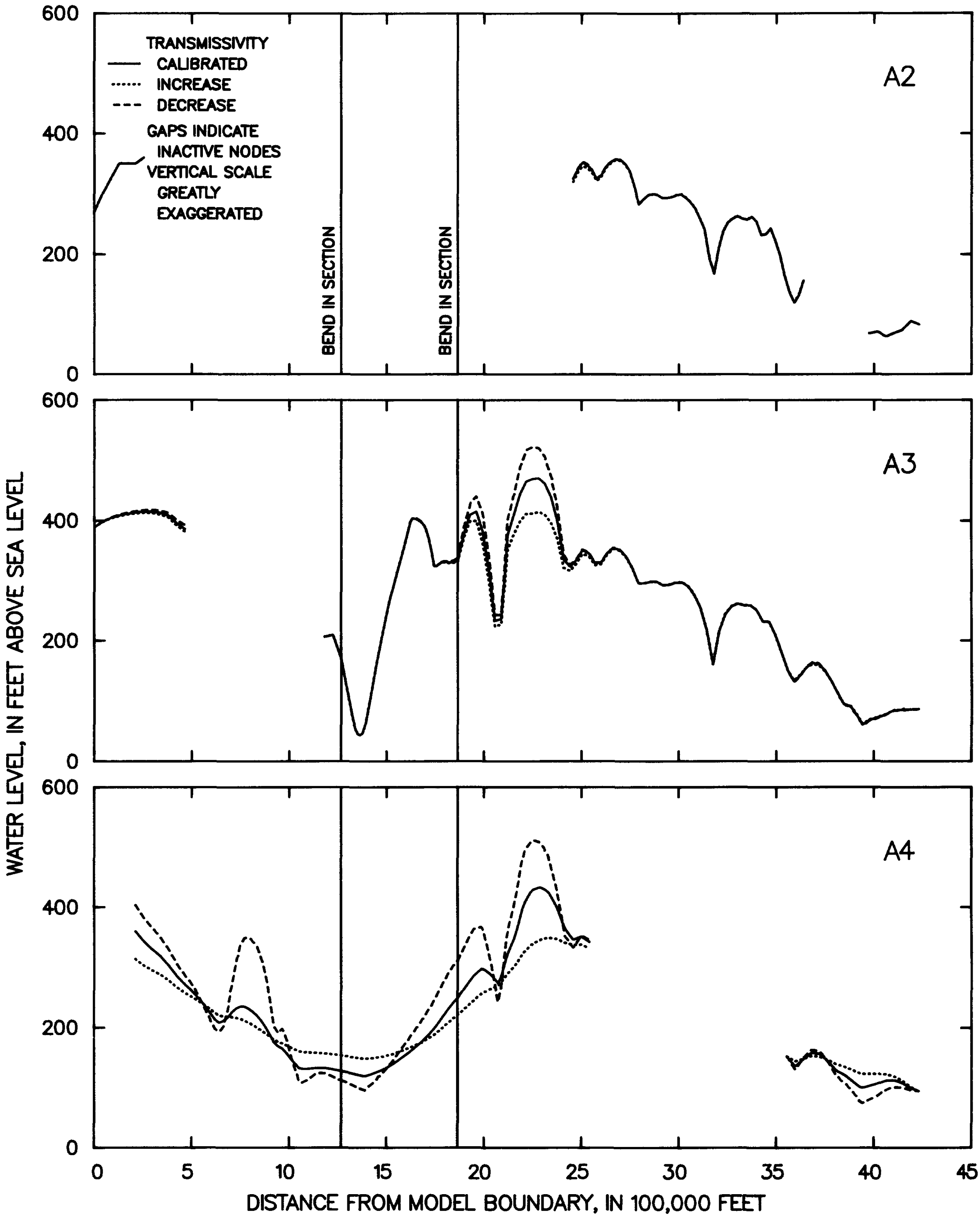

Figure 23.—Transverse head profiles from the calibrated A4 transmissivity and for a half an order of magnitude increase and decrease in A4 transmissivity.

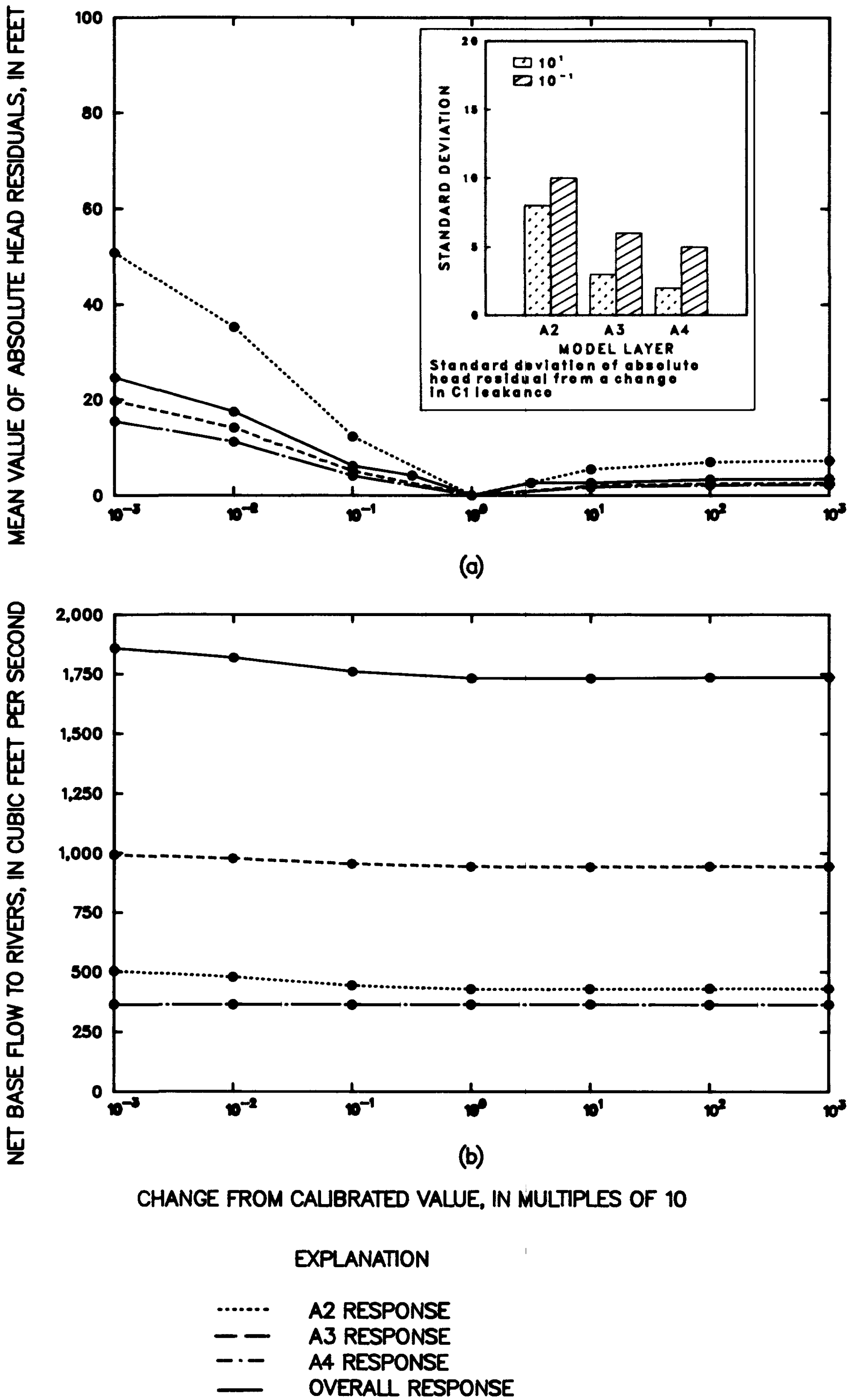

Figure 24.—Model sensitivity to changes in C1 leakance with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers .

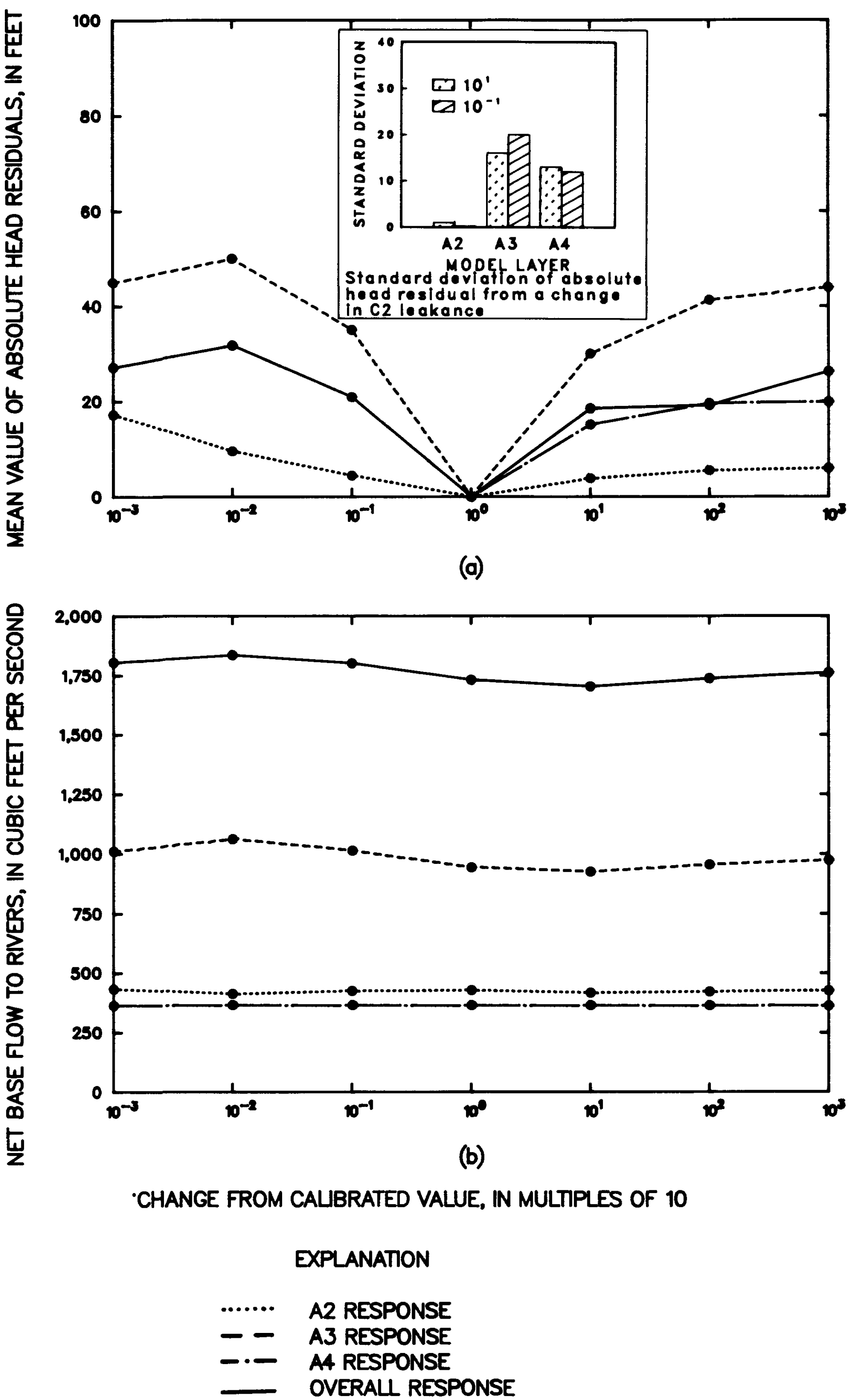

Figure 25.—Model sensitivity to change in C2 leakance with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

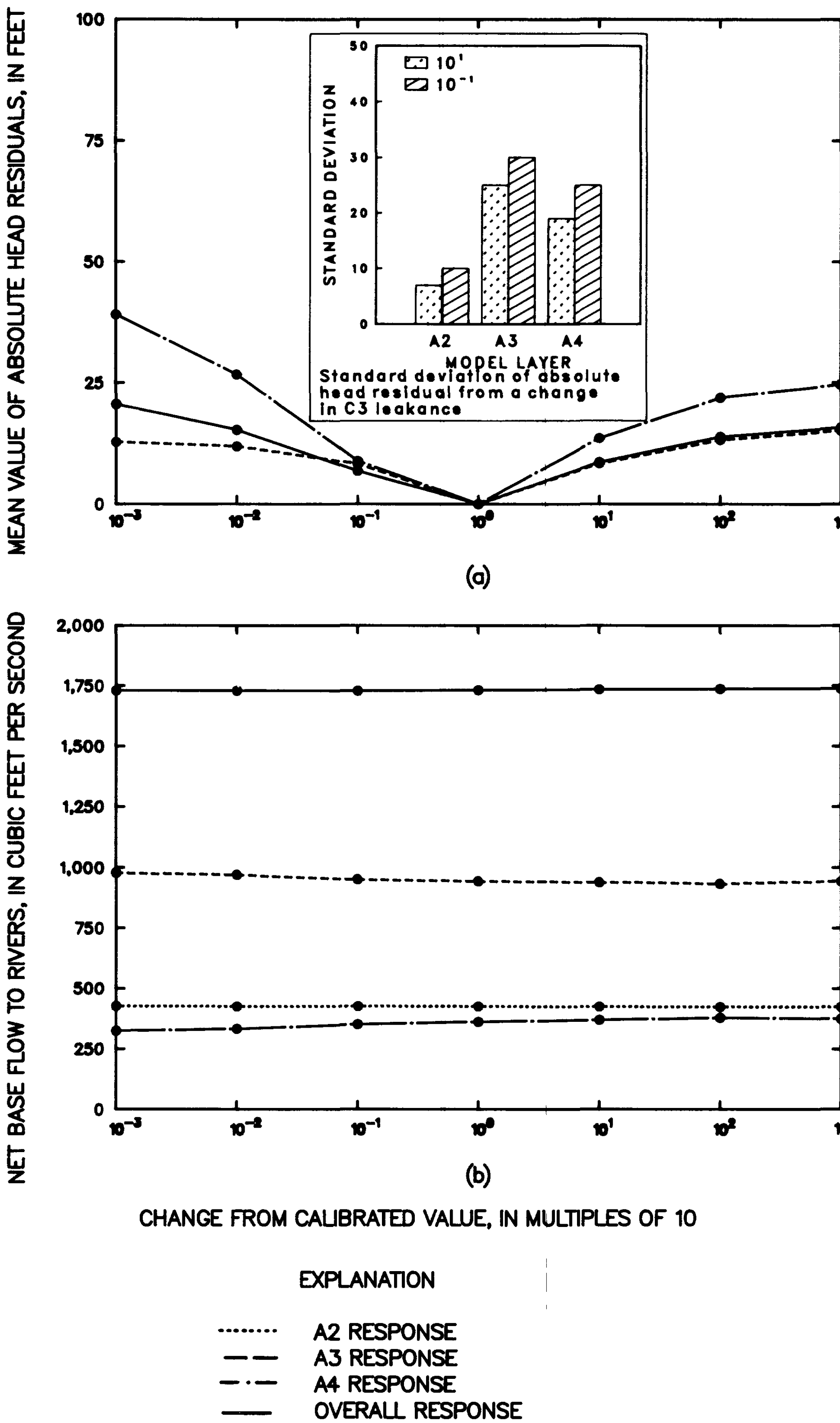

Figure 26.—Model sensitivity to change in C3 leakance with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers.

elevation of the area in question with respect to upgradient areas of recharge, and downgradient areas of discharge. Areas of high gradient are associated with topographically high outcrop areas, which receive recharge and leak water down to the lower layers. Low relief areas are generally downdip and associated with the confined part of the system that discharges water by diffuse upward leakage.

A change in the leakance of the C1 confining unit has the greatest effect on the heads in model layer A2, as illustrated in the longitudinal profile in figure 27. As leakance decreases due to the reduced conductivity of the conlayer A2 increase. The decrease in the C1 confining unit leakance produces an increase in model layer A3 and A4 heads, as the model responds to a new equilibrium with respect to the hampered upward leakage across the "tighter" C1 confining unit. Changes in the leakance of confining unit C1 have an insignificant effect on the heads in model layer A3 and A4 west of profile C-C′ (fig. 27) due to an increase in the thickness of the confining units separating these layers towards the west, which inhibits hydraulic communication between the aquifers.

The transverse head profiles (fig. 28) resulting from a change in the leakance of the C1 confining unit indicates only a slight response between the Flint and Savannah Rivers, as only a small area of contact exists between model layers SS and A2 at the location of this profile.

Longitudinal and transverse head profiles illustrating the simulated response to a half an order of magnitude change in the leakance of confining unit C2 are shown in figures 29 and 30, respectively. Changes in the leakance of this confining unit have the most impact on the heads in model layers A3 and A4 in Georgia and South Carolina (fig. 29). Under the calibrated status, discharge across the C2 confining unit allows model layer A3 to leak water upward to the overlying model layer A2 in the downdip areas. Under the scenerio of decreased leakance, the downdip heads in model layer A3 increased which, in turn, caused an increase in the updip heads in this layer. This increase in the updip heads in model layer A3 results because water that discharges downdip when using the calibrated leakance in confining unit C2 does not occur; the net effect is an increase in heads everywhere in model layer A3. A similar condition is illustrated in model layer A4 along the longitudinal head profile (fig. 29), as the model adjusts to the altered vertical gradient between the A3 and A4 aquifers.

The longitudinal and transverse head profiles resulting from a change in the leakance of confining unit C3 are shown in figures 31 and 32. The greatest response to a change in leakance in this confining unit occurs in the heads of model layer A4, except in Mississippi (profile A-A′, fig. 31). In general, a decrease in the leakance of confining unit C3 results in an increase in the downdip heads of model layer A3, because this area receives water that under calibrated conditions leaks down to the updip areas of model layer A4. In model layer A4, the heads decrease as less water is available via downward leakage from the updip areas of model layer A3. In Mississippi, confining unit C3 is relatively thick, with leakance values as small as 10^{-14} 1/s. Here, the response to a change in the leakance is mainly evident in model layer A3, with only a slight response in the profile of model layer A4.

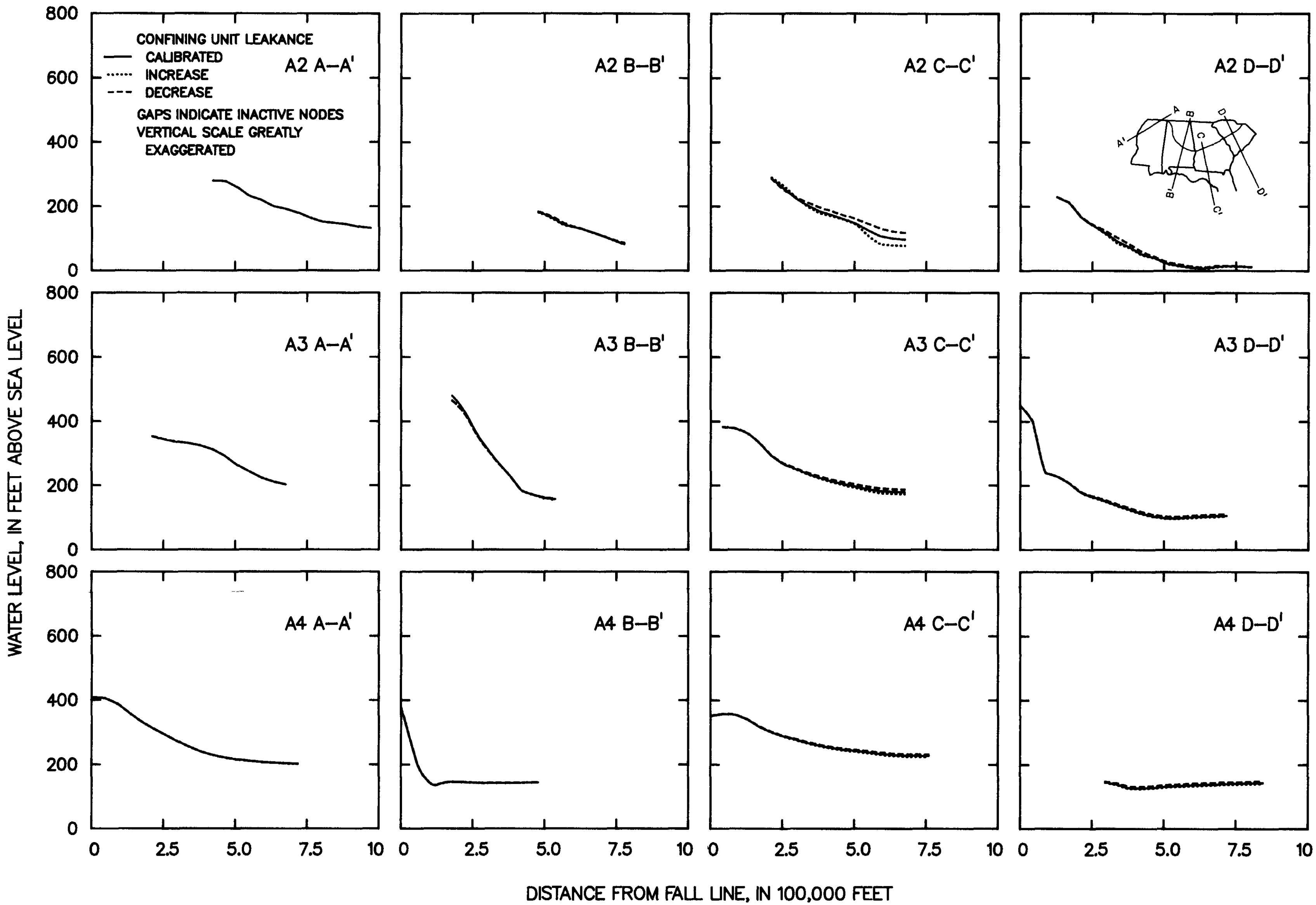

Figure 27.—Longitudinal head profiles from the calibrated C1 leakance and for a half an order of magnitude increase and decrease in C1 leakance.

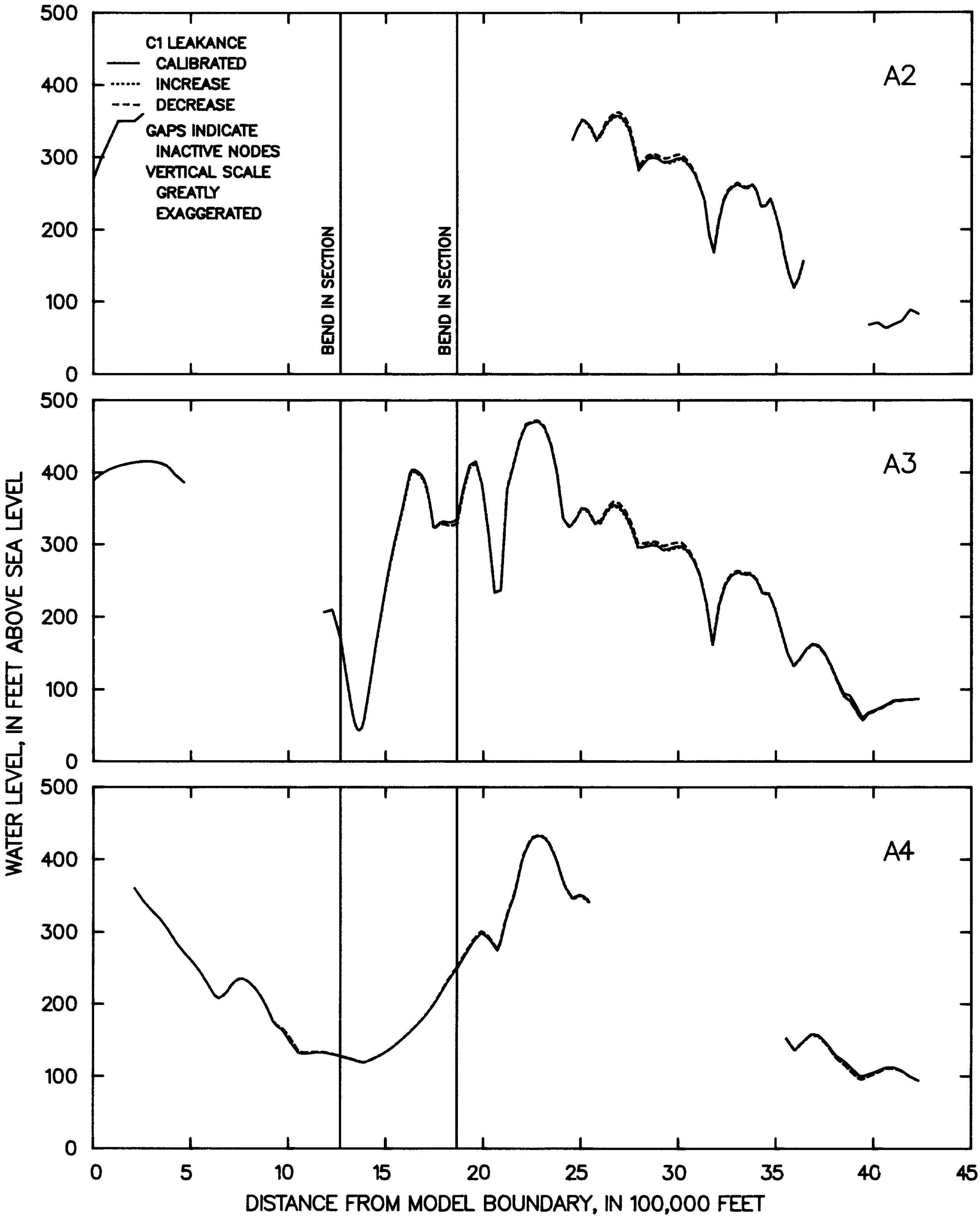

Figure 28.—Transverse head profiles from the calibrated C1 leakance and for a half an order of magnitude increase and decrease in C1 leakance.

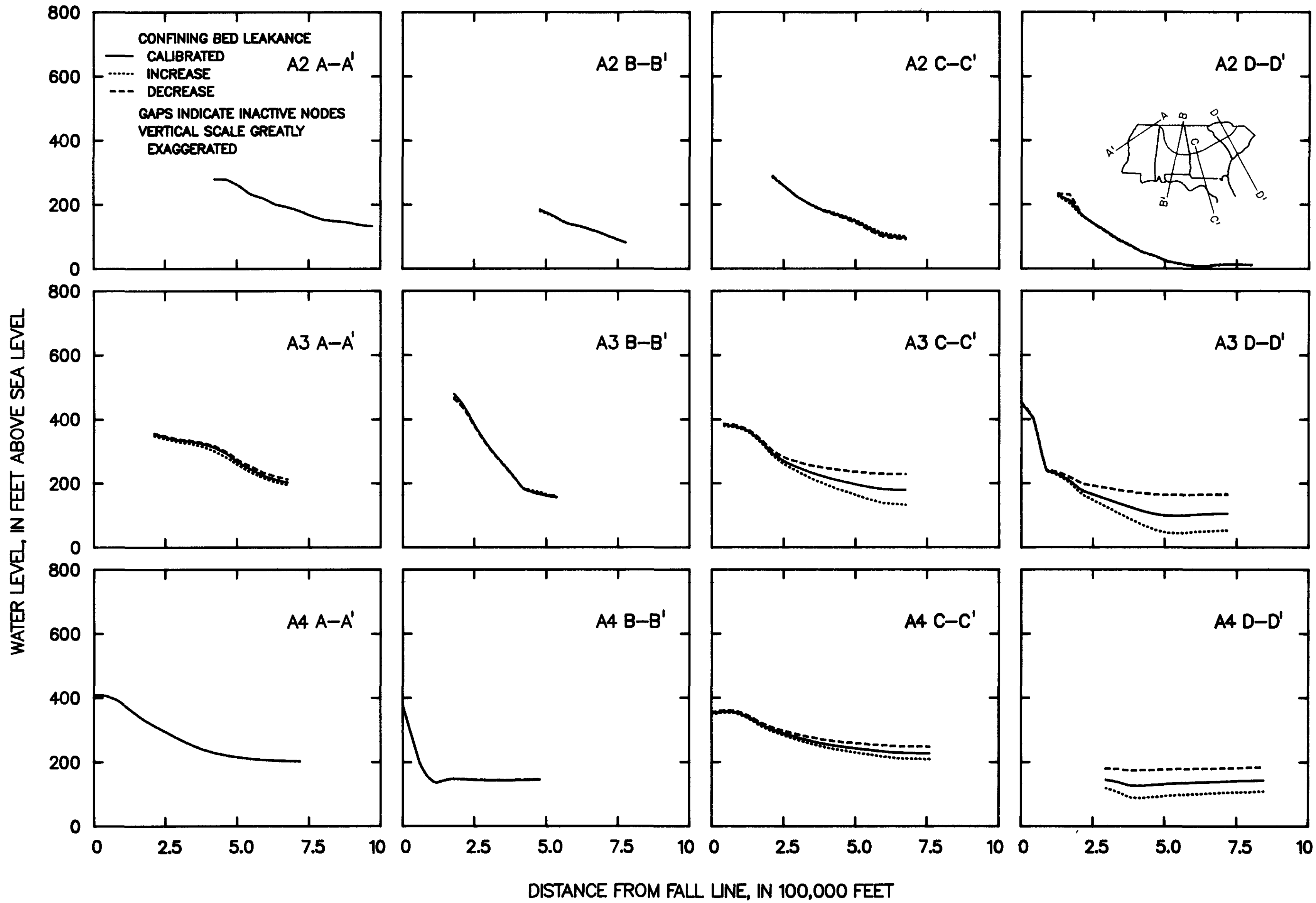

Figure 29.—Longitudinal head profiles from the calibrated C2 leakance and for a half an order of magnitude increase and decrease in C2 leakance.

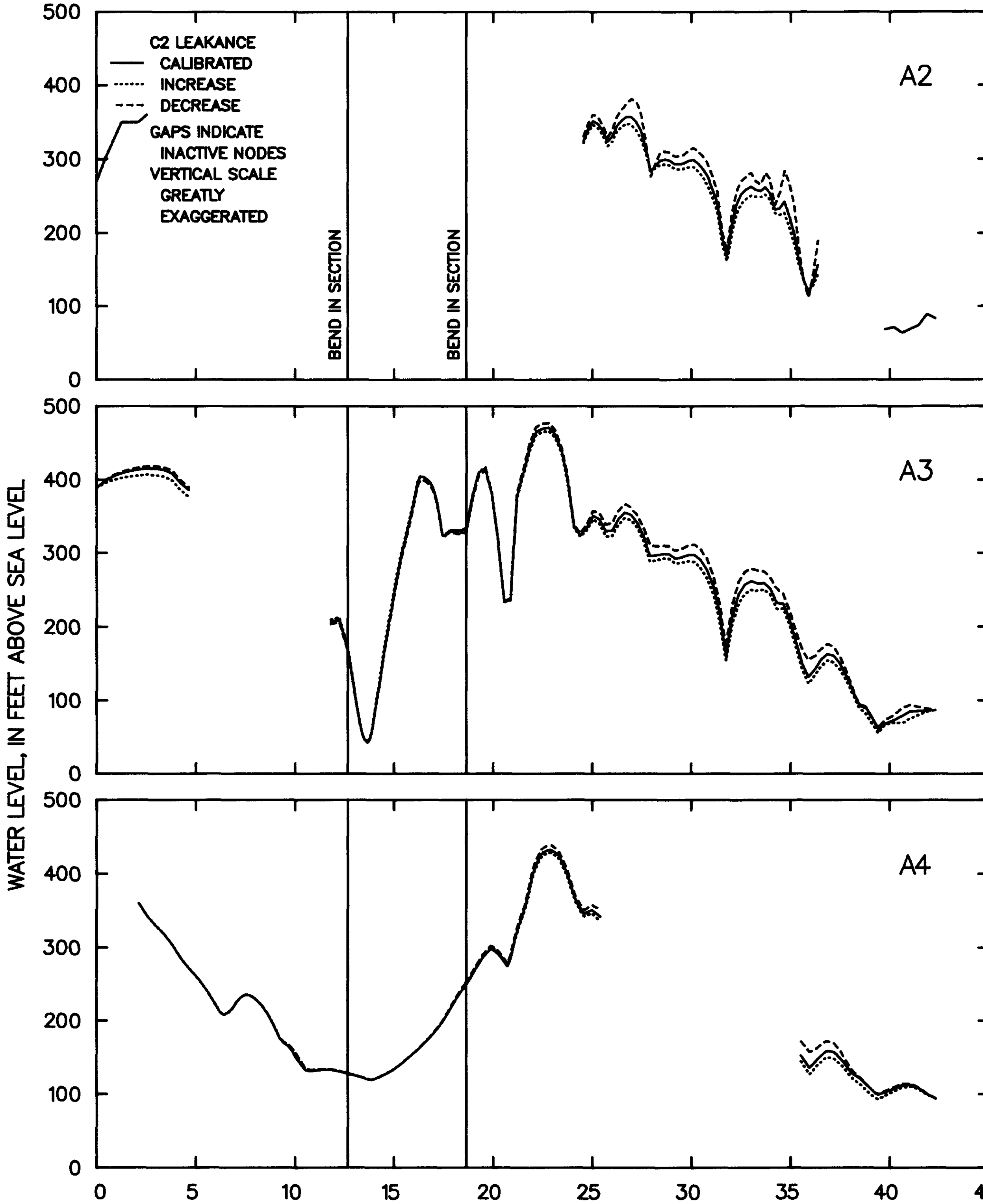

Figure 30.—Transverse head profiles from the calibrated C2 leakance and for a half an order of magnitude increase and decrease in C2 leakance.

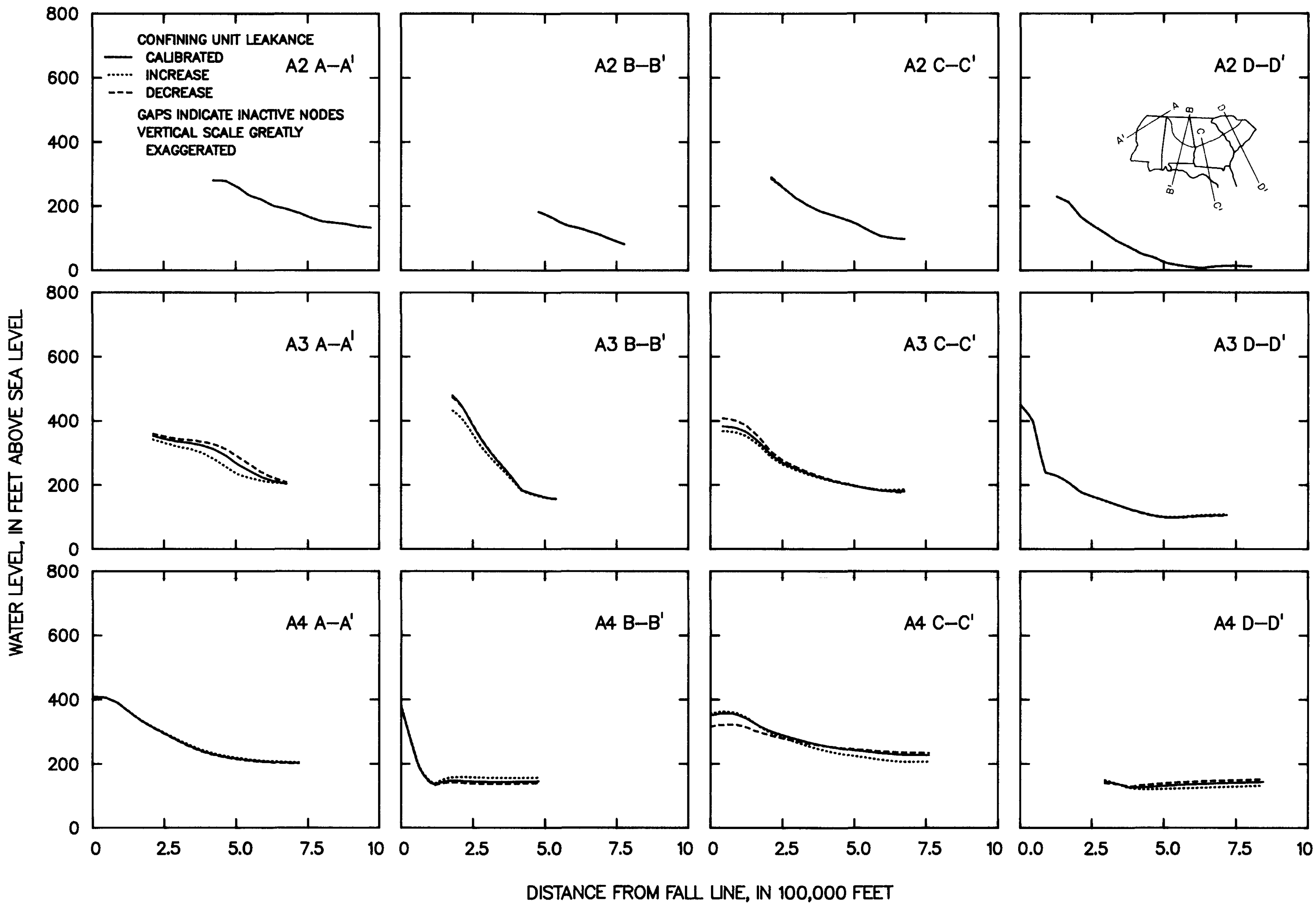

Figure 31.—Longitudinal head profiles from the calibrated C3 leakance and for a half an order of magnitude increase and decrease in C3 leakance.

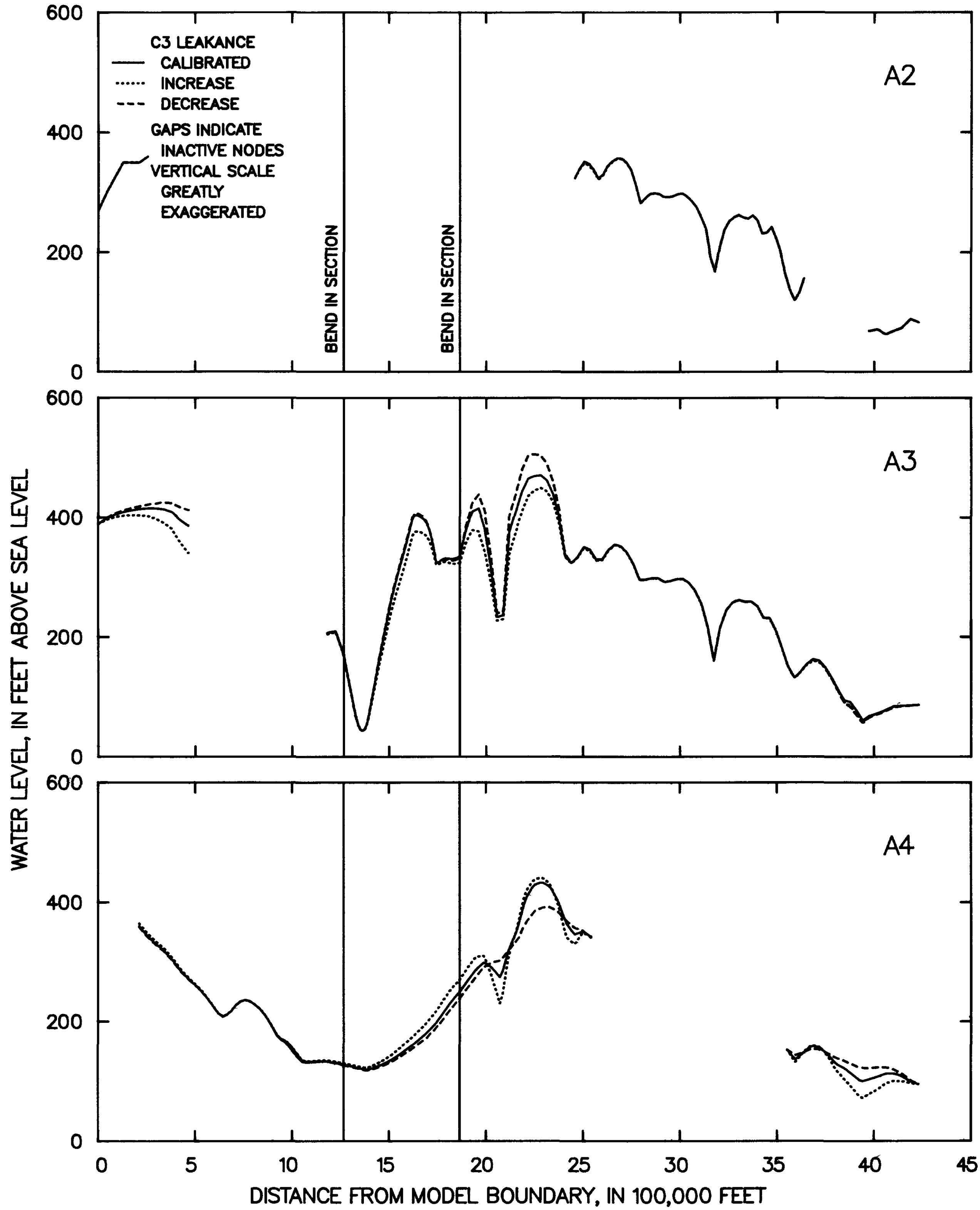

Figure 32.—Transverse head profiles from the calibrated C3 leakance and for a half an order of magnitude increase and decrease in C3 leakance.

In the updip areas of western Georgia (profile C-C´, fig. 31), confining unit C3 is thinnest (approximately 100 ft thick); as leakance decreases, the vertical flow between aquifers is reduced. This has a greater effect on lowering the heads in model layer A4 than it does on raising the heads in model layer A3, because transmissivity values are generally higher in model layer A3 than A4, thereby inducing lateral movement of the additional water to the rivers.

The head profile of model layer A4 in South Carolina (A4 D-D´, fig. 31) is unique to the system as it is isolated from both direct recharge and rivers. Lateral movement within this aquifer is minimal, as the transmissivities in this area average less than 4,000 ft^2/d and lateral gradients are relatively subtle; the principal means of discharge is upward leakage. A change in leakance of confining unit C3 has a greater effect on the head profile of model layer A4 than on the head in model layer A3 because layer A3 is connected to rivers which directly drain the aquifer system, whereas model layer A4 is comparatively isolated from the surficial drainage network in South Carolina.

Boundaries

The sensitivity of the model to changes in boundary conditions was limited to model layer A3 as the downdip boundary was based on a qualitative interpretation of mapped limits of hydraulic conductivity, whereas the boundaries in the other layers was based on a combination of hydraulic conductivity and geochemical information. The location of the downdip no-flow boundary and the altitude of the constant head boundaries was varied. The results are shown in figures 33 and 34, respectively. The no-flow boundary was moved uniformly in the downgradient direction to the edge of the model grid (equal to 4 model nodes or 32 mi). The sensitivity of moving the boundary in 4 model nodes was not tested, as a no-flow boundary was considered inappropriate in this area, and it was not the intent of this analysis to experiment with other boundary conditions. The altitude of the constant boundaries were increased and decreased 10 ft.

The head change resulting from moving the boundary and from changing the constant head altitude has the greatest impact on the heads near the shift in boundary conditions, and the head change decreases rapidly with distance from the boundary. The area of influence of moving the no-flow boundary was about 10 nodes (or 80 mi) with the average head change in this area being about 2 ft. A change in the constant head altitude increased a greater area along the northwest boundary (located in Mississippi) than it did along the northeast boundary (located in South Carolina), but in both areas the average head change was less than 5 ft.

The model appears to be relatively insensitive to moderate adjustments in the boundary conditions of model layer A3. The differences in model area affected by the boundary change appears to be more of a function of the differences between the magnitude of the adjacent transmissivities than it is to a change in the lateral gradients.

MODEL SENSITIVITY TO TRANSIENT-STATE INPUT

The sensitivity analysis of the transient-state model was limited to changing the storage coefficient, as this was the only variable adjusted in the transient calibration process. The simulated results were recorded as a change

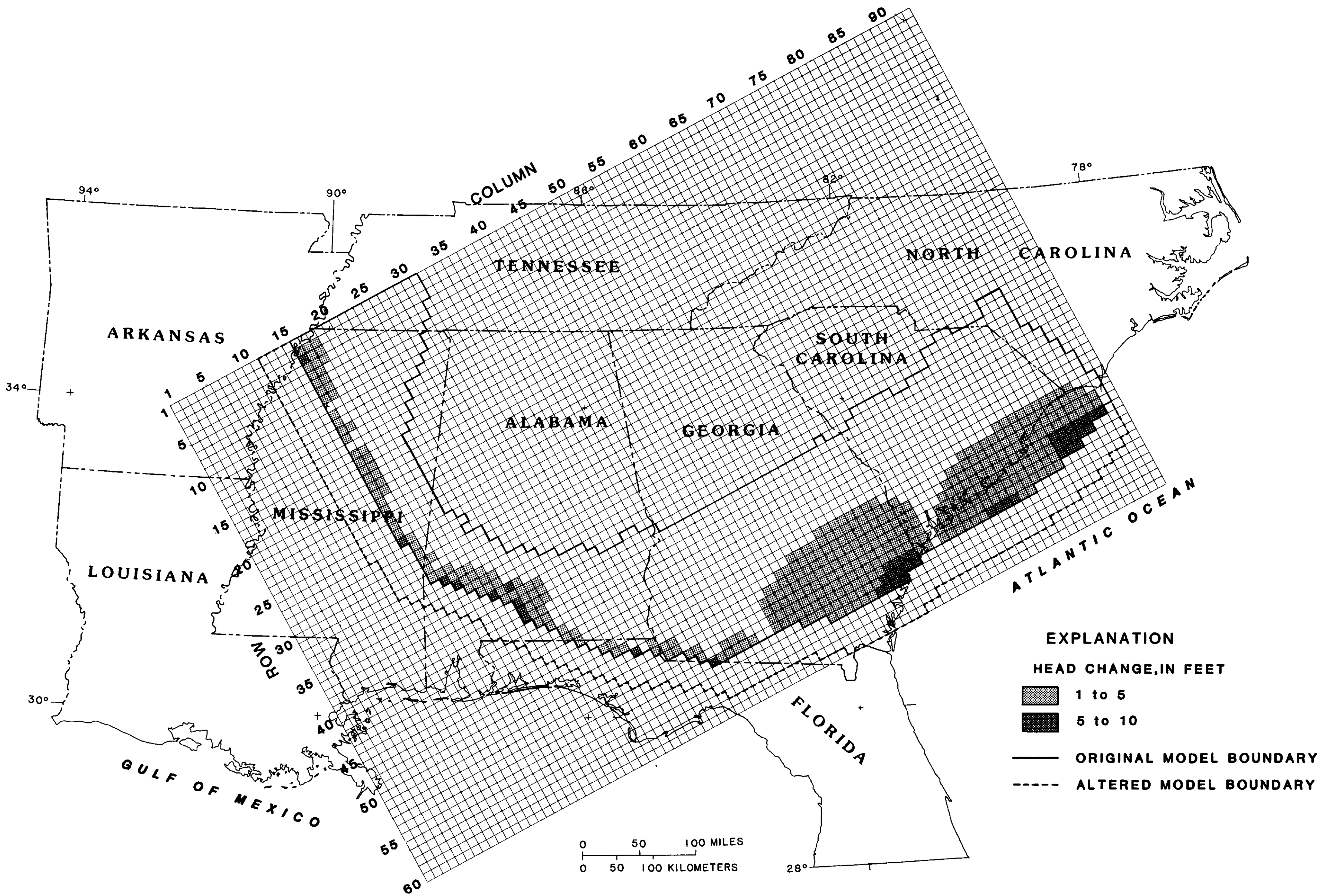

Figure 33.—Area and magnitude of head change resulting from moving the no-flow boundary in model layer A3 downdip 32 miles.

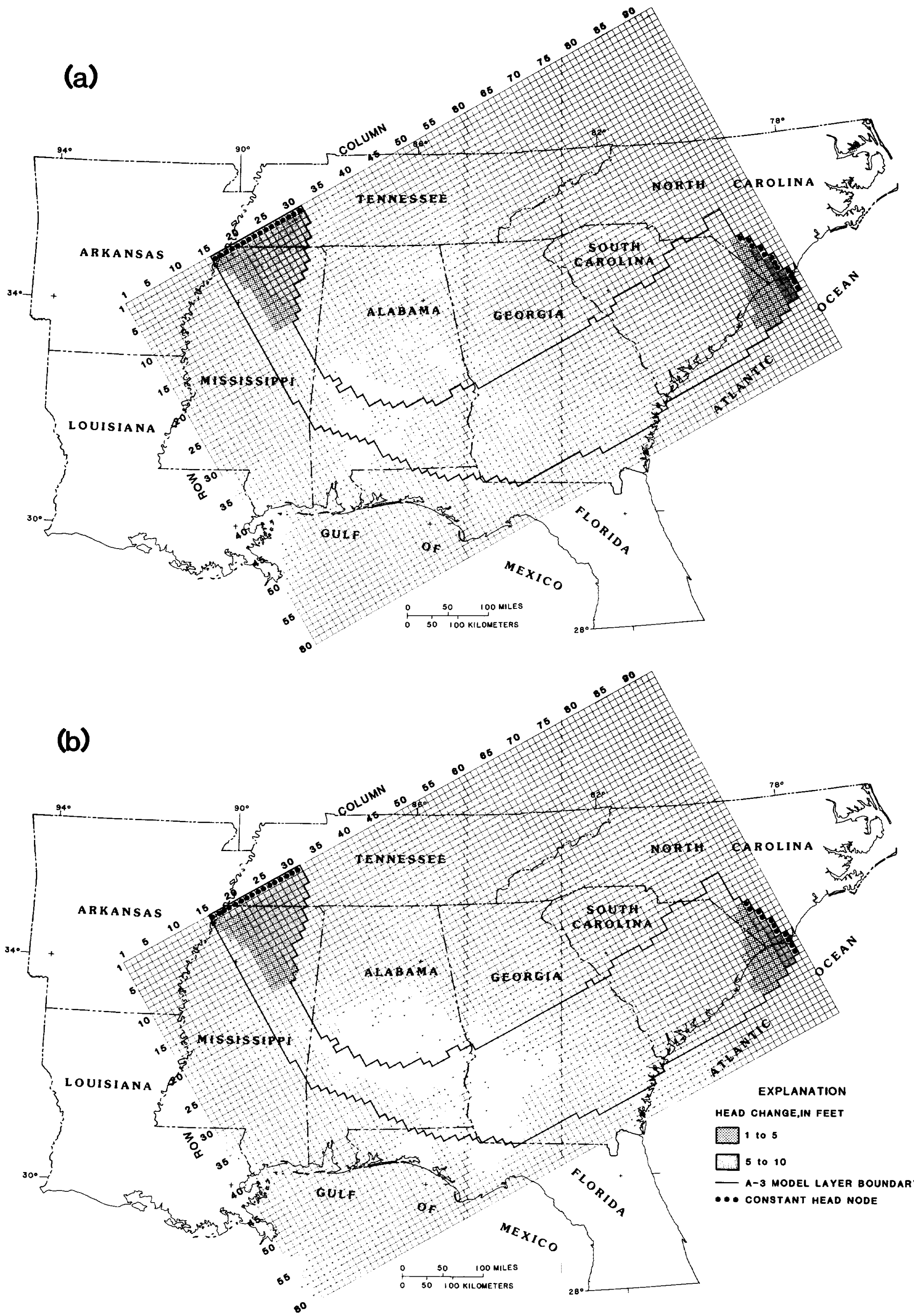

Figure 34.—Area and magnitude of head change resulting from (a) a 10-foot increase and (b) a 10-foot decrease in elevation of the A3 constant head boundary.

in the absolute head residual averaged in all layers and the sum of the base flow to rivers. In the transient model, the head residual is a measure of drawdown. The standard deviation of the absolute head residual was calculated to evaluate the relative sensitivity of the model to changes in the storage values. Hydrographs showing major pumping centers and widespread areas of drawdown were used to illustrate the decline in water levels in model layers A3 and A4 resulting from a change in storage coefficient. The location of these major pumping centers is shown in figure 35.

Storage Coefficient

The storage coefficients were uniformly increased or decreased in all layers simultaneously up to an order of magnitude from the calibrated values so that the changed values would range from 1×10^{-02} to 5×10^{-05}. The higher values represent semiconfined conditions in the outcrop areas, while the smaller values represent confined conditions in mid and downdip areas of the system. The results of the sensitivity analysis are shown in figure 36, where base flow is shown to decrease over time, and the head residual, or drawdown, increases over time.

When storage coefficient is decreased relative to calibrated conditions, less water is released from aquifer storage; drawdown increases above the calibrated profile, and base flow decreases as the gradients towards the streams flatten, causing the river heads to drop below the stage. When storage coefficient are increased, more water is removed from storage, there is less drawdown, and the net base flow to the rivers increases as gradients are steeper.

Simulated hydrographs were compared to six hydrographs drawn from observed head data to illustrate the influence of alternative storage coefficient values on the simulation of heads (fig. 37). Lowering the storage coefficient produced lower simulated heads which in some cases appears to reduce the divergence between simulated and observed hydrographs. However, storage coefficients less than those assumed for calibration are not compatible with results of aquifer tests. Because lower storage coefficient values approach the compressibility of water (10^{-6}), they are not considered physically realistic, nor appropriate for simulation of regional aspects of Coastal Plain hydrogeology.

SUMMARY AND CONCLUSIONS

A sensitivity analysis was conducted on a multilayer ground-water flow model to determine the effect that changes in the values of input parameters have on simulated water levels and base flow to rivers. The input parameters were uniformly increased and decreased by as much as three orders of magnitude in an attempt to reach the model's threshold where further changes have little or no effect on the simulation of base flow and hydraulic head. The simulated water levels were recorded statistically as the mean value and standard deviation of the absolute head residual. The results were illustrated as hydraulic head profiles drawn longitudinal and transverse to the general flow direction. Net base flow was computed as the algebraic difference between the aquifer-to-stream flow and the stream-to-aquifer flow, although the latter flow was usually insignificant.

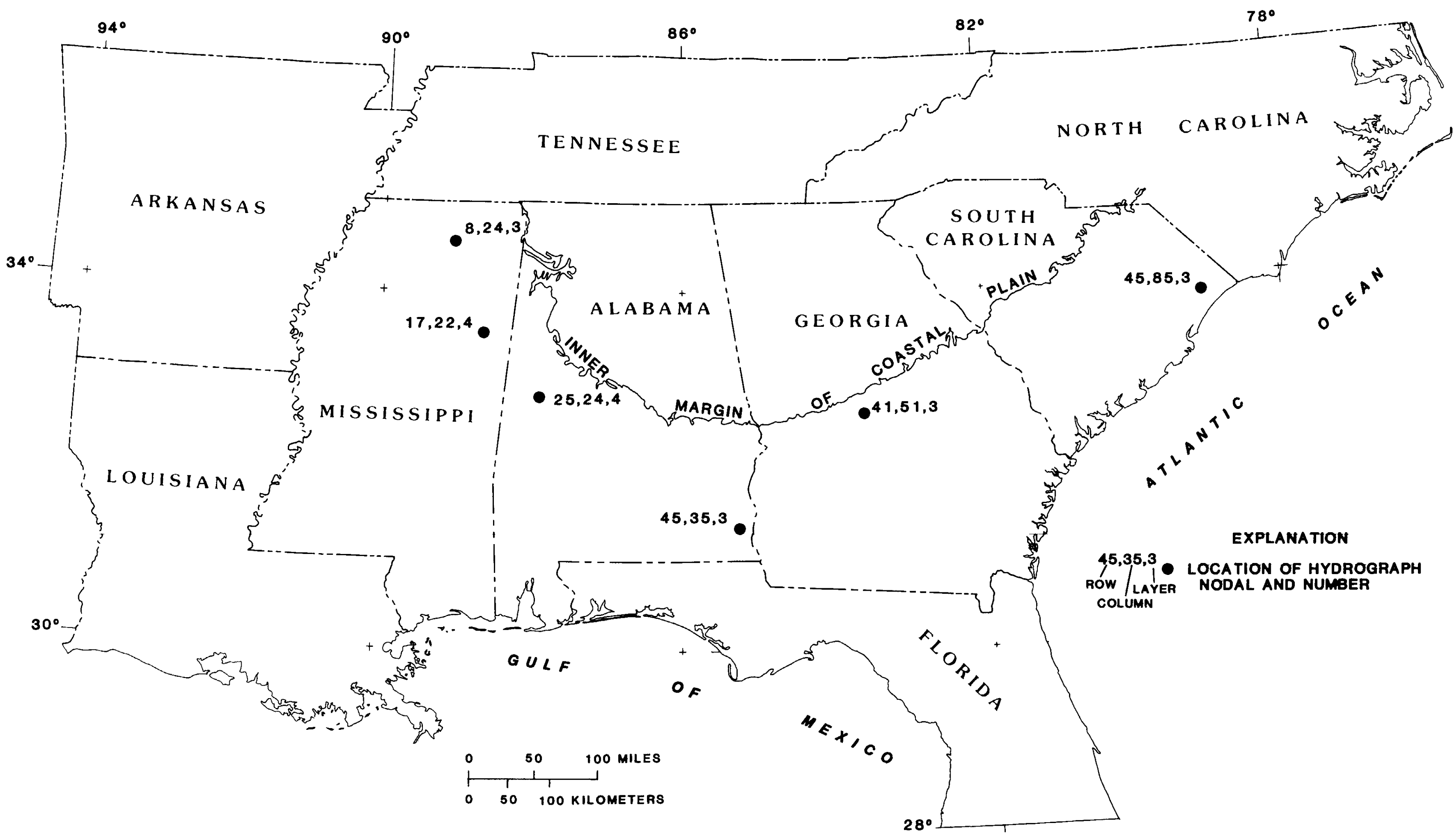

Figure 35.—Observed hydrographs at selected sites in the Southeastern Coastal Plain aquifer system.

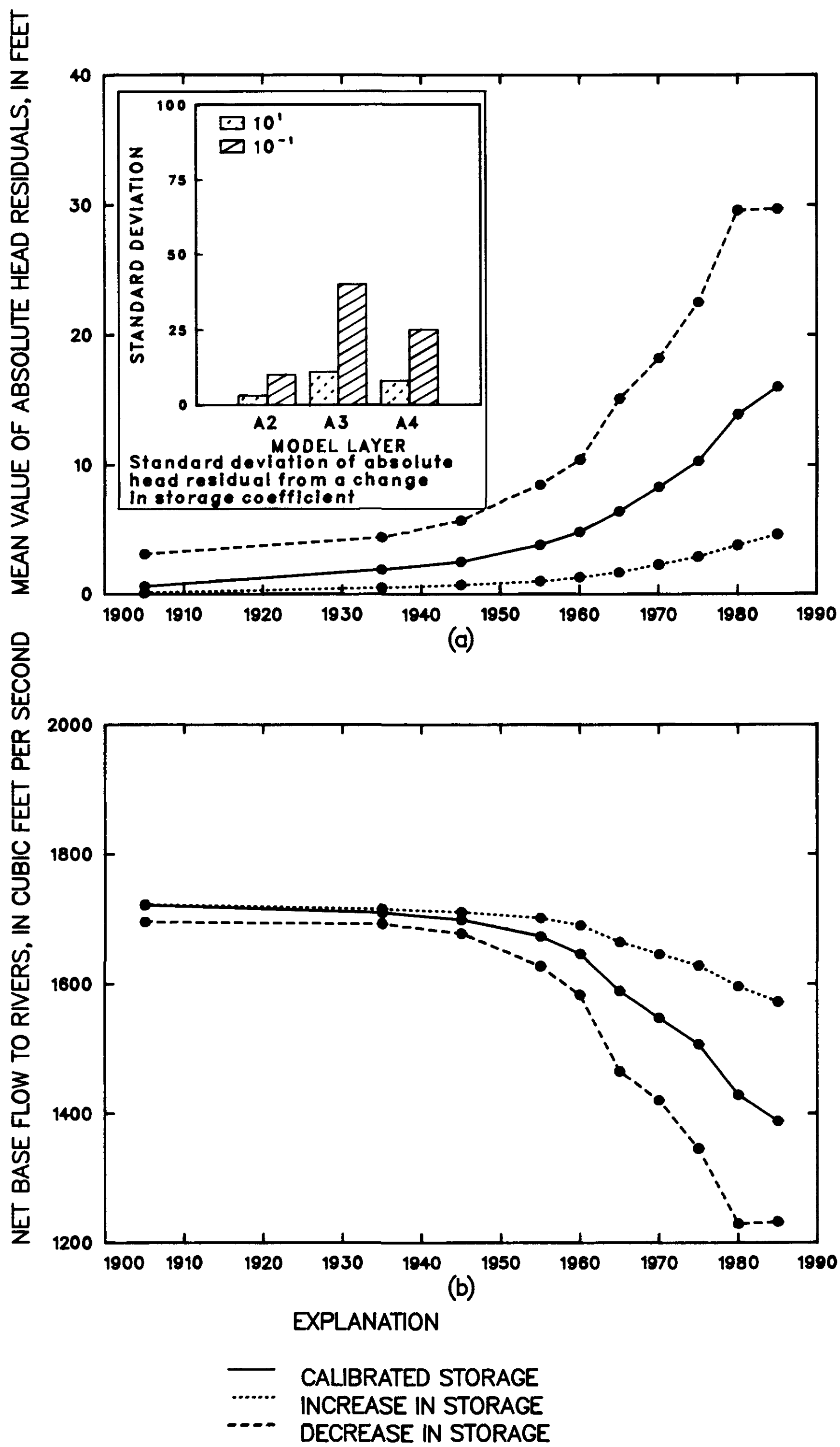

Figure 36.–Overall model response to change in storage coefficient with respect to (a) mean value and standard deviation of absolute head residuals, and (b) net base flow to rivers from the calibrated model, and from an order of magnitude increase and decrease in storage coefficient.

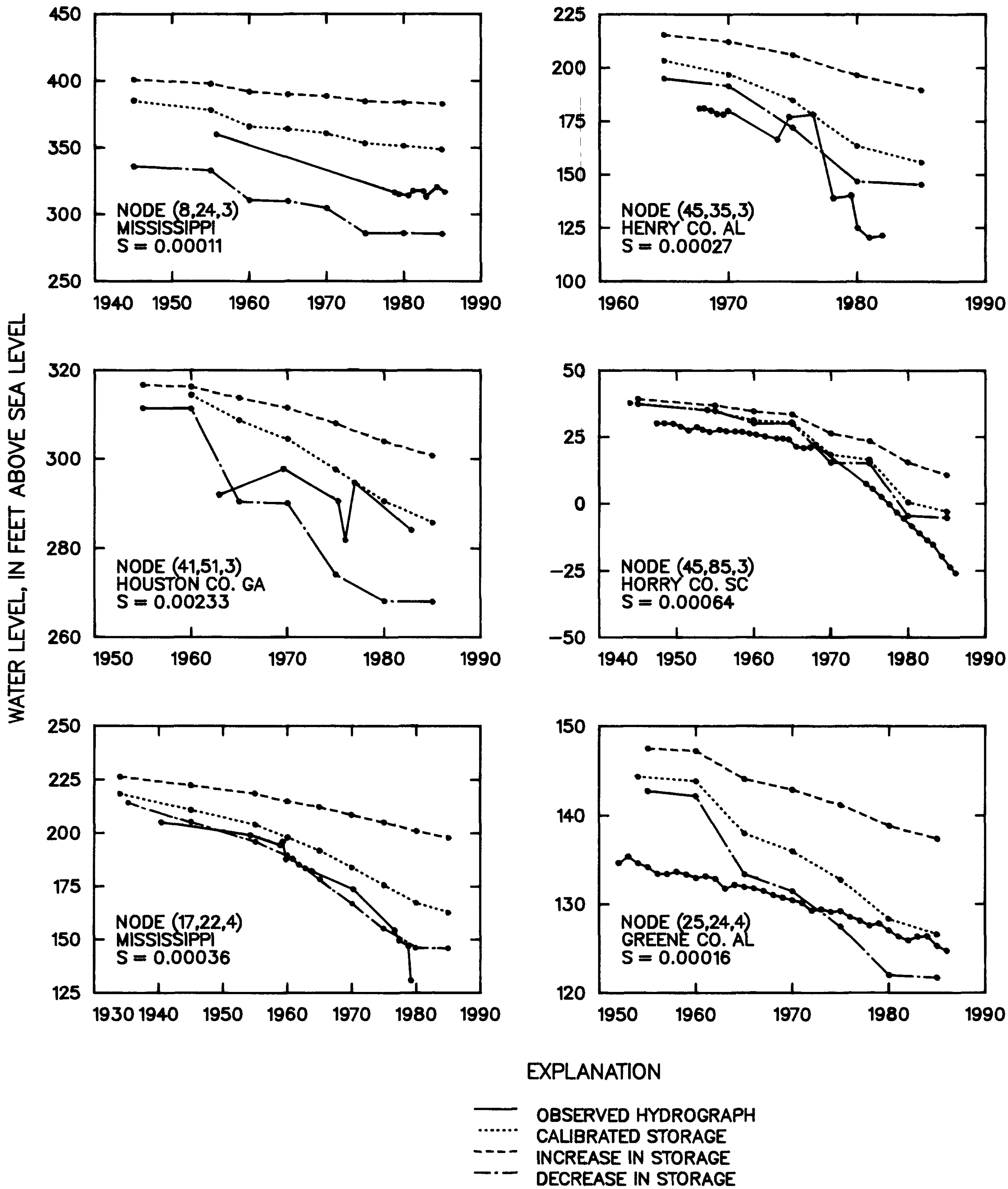

Figure 37.—Decline in water level at selected sites, from the calibrated storage coefficient, and for an order of magnitude increase and decrease in storage.

The tested parameters were limited to those routinely adjusted during calibration with the intent to simulate published predevelopment potentiometric surfaces and estimated rates of base flow. Parameters tested in the steady-state model were recharge, riverbed conductance, transmissivity, confining unit leakance, and selected boundary conditions. In the transient-state model, only changes in the value of storage coefficient were tested.

The sensitivity of each parameter was tested independently of the others to assess the relative importance of each parameter on calibration. This is not to imply that the input parameters act independently of each other. In fact, the simulated response is often due to a combination of factors. For example, when recharge is changed, both the transmissivity in the area where the recharge is applied and the riverbed conductance values are important in determining whether the additional recharge will result in a build up of head or whether the additional recharge will discharge to the streams. In model layers A3 and A4, the transmissivity distributions near the outcrop of each model layer are very similar (values in both layers are greater than 8,000 ft^2/d), but the riverbed conductance values of the rivers in model layer A3 are 3 to 4 times higher than those in model layer A4. As a result, when recharge increases, the additional recharge discharges to the rivers in model layer A3, whereas in model layer A4 the additional recharge causes an increase in hydraulic head.

The steady-state model was shown to be most sensitive overall to increases in the rates of recharge. The standard deviation resulting from an order of magnitude increase in recharge averaged nearly 500 ft in all model layers, whereas a comparable change in riverbed conductance resulted in a standard deviation of less than 150 ft. When the recharge rate is held constant, the model is more sensitive to variations in transmissivity, especially updip in the interstream divide areas, where the hydraulic gradients are steep. Near the rivers, the model is more sensitive to changes in riverbed conductance than it is to comparable changes in transmissivity and confining unit leakance. The model is relatively insensitive to confining unit leakance in terms of affecting base flow to the rivers, but is very sensitive with respect to the simulation of head distributions, especially in areas where confining units are thin or the vertical hydraulic conductivity of the confining unit is large. The model is relatively insensitive to moderate changes in the location of the no-flow boundary in model layer A3, and to the altitude at the constant head boundaries of this layer.

The transient-state model is more sensitive to a decrease in storage coefficient than to an increase. As storage coefficient is decreased, less water is released from storage and, as a result, the head residual (or in this instance, the drawdown) increases above the calibrated profile. Increased drawdown decreases the head gradients towards the rivers, and in some cases causes the head in the river to drop below the river stage, resulting in a reduction of base flow compared to calibrated conditions.

Simulated base flow and simulated head residual were plotted against the order of magnitude change in the value of the altered parameter. The resulting response curves and the associated standard deviation of the absolute head residual were used to evaluate the appropriateness of the calibrated values in terms of accurately representing the physical system. When the calibrated value falls on the steep part of the response curve (such as the case for recharge), a relatively high degree of confidence in the calibrated value is probably justified, as a change from the calibrated status would generally result in significant departures from observed conditions. When the calibrated value falls on the flat part of the response curve (such as the case for confining unit leakance), the model is relatively insensitive to that parameter, providing little or no indication of how much confidence should be placed in the calibrated value in regards to how well it approximates conditions in the real system.

The standard deviation of the absolute head residual tended to be 3 to 6 times greater when a given input parameter was decreased, compared to when the value of the parameter was increased by the same amount. The exception to this is the case for recharge, in which the model is far more sensitive to an increase than it is to a decrease, with the difference in the standard deviation being a factor of 10. In terms of head residuals, the model is generally less sensitive to an increase in a given input parameter than it is to a decrease.

The overall results of the sensitivity analysis suggest that the model's ability to simulate actual conditions deteriorates as departure from the calibrated values of the input parameters increase. In general, the calibrated values of the model parameters tested are centrally clustered within the range of reasonable values that might be used to represent the physical system. In the updip and middip areas of the model, where data was generally available with which to calibrate the model, there is a higher degree of confidence in the calibrated model parameters than in the downdip areas of the model, where comparatively little is known about the flow system.

REFERENCES

Barker, R. A. 1986, Preliminary results of a steady-state ground-water flow model of the Southeastern Coastal Plain aquifer system, In Proceedings of the Southern Regional Ground Water Conference, September 18-19, 1985, San Antonio, Texas: The Association of Ground Water Scientists and Engineers, Division of the National Water Well Association, p. 315-338.

Bush, P. W. 1982, Predevelopment flow in the Tertiary limestone aquifer, Southeastern United States: a regional analysis from digital modeling: U.S. Geological Survey Water-Resources Investigations Report 82-905, 41 p.

McDonald, G. W. and Harbaugh, A. W., 1984, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 83-875, 528 p.

Miller, J. A., 1984, Hydrogeologic framework of the Floridan aquifer system in Florida and parts of Georgia, South Carolina, and Alabama: U.S. Geological Survey Professional Paper 1403-B, 91 p.

Renken, R. A., 1984, The hydrogeologic framework for the Southeastern Coastal Plain aquifer system of the United States: U.S. Geological Survey Water-Resources Investigations 82-4243, 26 p.

Stricker, V. A., 1984, Approximate potentiometric surface for aquifer unit A4, Southeastern Coastal Plain aquifer system, of the United States, prior to development: U.S. Geological Survey Water-Resources Investigations Report 84-4364, scale 1:2,000,000, 1 sheet.

_____ 1985a, Approximate potentiometric surface for aquifer unit A2, Southeastern Coastal Plain aquifer system, of the United States, prior to development: U.S. Geological Survey Water-Resources Investigations Report 85-4019, scale 1:2,000,000, 1 sheet.

_____ 1985b, Approximate potentiometric surface for aquifer unit A3, Southeastern Coastal Plain aquifer system, of the United States, prior to development: U.S. Geological Survey Water-Resources Investigations Report 85-4031, scale 1:2,000,000, 1 sheet.

Wait, R. L. and Davis, M. E., 1986, Configuration and hydrology of the Pre-Cretaceous rocks underlying the Southeastern Coastal Plain aquifer system: U.S. Geological Survey Water-Resources Investigations Report 86-4010, 1 sheet.

Wait, R. L., Renken, R. A., Barker, R. A., Lee, R. W., and Stricker, V. A., 1986, Southeastern Coastal Plain Regional Aquifer-System Study, p. 205-222, In, Sun, R. J., editor, Regional Aquifer-System Analysis Program of the U.S. Geological Survey, Summary of Projects, 1978-84. U.S. Geological Survey Circular 1002, 264 p.