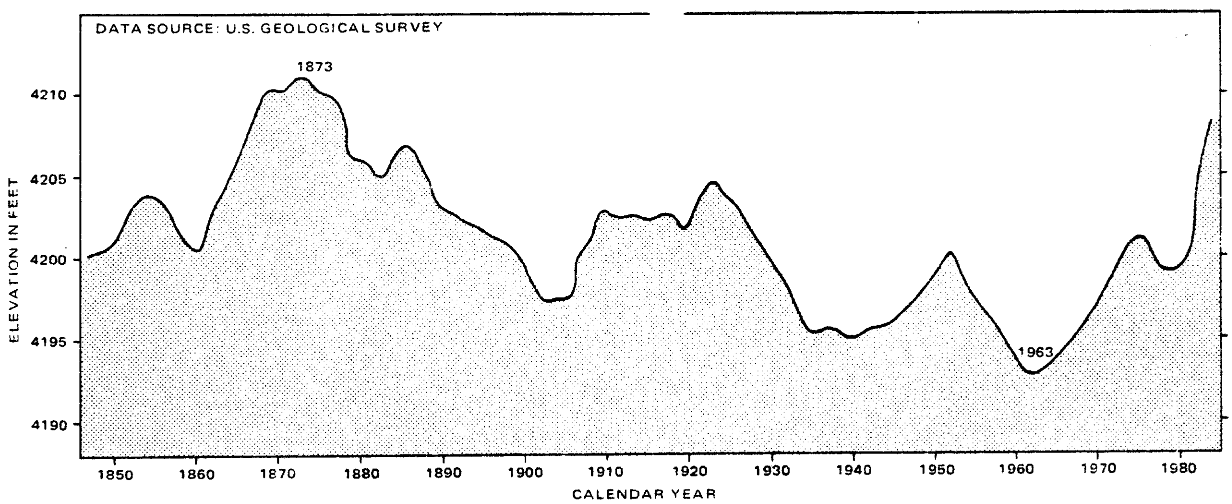
PROBLEMS OF AND PROSPECTS FOR PREDICTING GREAT SALT LAKE LEVELS

PAPERS FROM A CONFERENCE HELD IN SALT LAKE CITY
MARCH 26-28, 1985



EDITED BY

Paul A. Kay, University of Utah
Henry F. Diaz, NOAA/ERL

·nter for Public Affairs and Administration, University of Utah
May 1985

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National Oceanic and Atmospheric Administration
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INTRODUCTION

This publication had its origin in discussions between the two editors in late summer of 1984 concerning the historically unprecedented wetness of the preceding two years and the rapid rise of Great Salt Lake levels. We felt that the time was ripe for a meeting of meteorologists, climatologists and hydrologists to conduct a scientific assessment of our understanding of the lake as a response to climate and of our ability to forecast lake behavior. At about the same time, and independent of our discussions, the Federal Emergency Management Agency's 30-day report on flood hazard mitigation recommended the convening of "a workshop to refine the interim working elevation of 4217 feet and offer a methodology to study the lake" (FEMA-720-DR-UTAH, September 14, 1984, p. 15). Discussions were also underway in the Utah Geological and Mineral Survey about the possibility of a meeting of geologists to address the late Quaternary record of lake levels. Accordingly, a joint planning meeting was held at the University of Utah in October 1984, from which emerged the outline and strategy for the conference that resulted in this publication.

The conference was held in Salt Lake City on 26-28 March 1985. Invited speakers presented reviews of the current status of their research into climatic fluctuations, hydrologic modelling, short and long term records of lake level fluctuations, and related topics. The theme was an assessment of our scientific knowledge of the lake system; the conference was not intended to debate or formulate policy positions. Our original suggetion was that the presentations review work already completed or in progress; to our delight, many participants presented work especially performed (on very short notice) for the conference. There was ample discussion of each presentation, and the final session was devoted to summaries by a panel who, although not directly working on the problem of Great Salt Lake were nevertheless expert in the type of questions raised (George I. Smith, USGS, and Roger G. Barry and Colin S. Ramage, CIRES). An open discussion among the participants, led by Genevieve Atwood, UGMS, reviewed the proceedings and sought consensus on a "planning level" or "level of concern" and on needs for further research. It was abundantly clear that there is still a great deal to learn about the lake and the climatic system that drives it; a series of recommendations arising from the conference are summarized at the end of this volume.

Given the immediacy of the interest in lake level questions and the needs of the planning and management community, we intended rapid publication of the conference proceedings. Therefore, we have foregone formal peer review of the papers. The papers have been only lightly edited, and no uniformity of style has been imposed. This volume contains all the papers presented at the meeting, with session summaries and summary of the final discussion. The authors have been encouraged to submit expanded reports for refereed publication elsewhere if they so desire.

That the conference ran as well as it did and that this publication appears so soon afterwards are due in no small part to the logistical support provided by the staffs of the Office of Conferences and Institutes, the Center for Public Affairs and Administration, and the Department of Geography, University of Utah. Our appreciation also goes to Clancy Philipsborn and Nancy Stone of FEMA, Region VIII, for their support and suggestions, particularly in running a very successful field trip to the south and east shores of the lake. Financial support for the conference was provided by the following agencies, to whom we all owe thanks for making the conference possible: College of Social and Behavioral Science, University of Utah; National Oceanic and Atmospheric Administration; Federal Highways Administration; Utah Department of Transport; and, the Office of the State Planning Coordinator (Utah).

Paul A. Kay
Henry F. Diaz

SESSION 1

PALEOENVIRONMENTAL RECORD OF LONG-TERM FLUCTUATIONS

Durations, Average Rates, and Probable Causes of Lake Bonneville Expansions, Stillstands, and Contractions During the Last Deep-Lake Cycle, 32,000 to 10,000 Years Ago

Donald R. Currey[1] and Charles G. Oviatt[2]

[1] Department of Geography, University of Utah, Salt Lake City, UT 84112
[2] Utah Geological and Mineral Survey, Salt Lake City, UT 84108

Introduction

Great Salt Lake has fluctuated widely at levels below 1300 m (4265 ft) during the Holocene Epoch of the last 10,000 years. Great Salt Lake's hydrologic basin of nearly 60,000 sq km (23,000 sq mi) occupies the northeastern 43 percent of the Lake Bonneville hydrologic basin. As an Ice Age predecessor of Great Salt Lake, Lake Bonneville fluctuated at levels as high as 1552 m (5092 ft) during the last deep-lake cycle of the Pleistocene Epoch, between 32,000 and 10,000 years ago. The full range of lacustrine variability that could conceivably occur in the region in the future is probably represented in the stratigraphic record of the last deep-lake cycle. The purpose of this paper is to extract from portions of that record some estimates regarding durations, average rates, and probable causes of Lake Bonneville expansions, stillstands, and contractions. These estimates are intended to place predictive studies of Great Salt Lake fluctuations in a broad context by providing a long-term, extreme-case perspective on the region's hydrologic potential.

The Lake Bonneville-Great Salt Lake closed-basin system is a climate-forced hydrologic system with a very large capacity to store information regarding its past history. That information can be retrieved, interpreted, and applied in a variety of ways (Figure 1). Vertical and lateral arrangements of sedimentary lithofacies, biofacies, and chemofacies, together with diagnostic properties of those facies, are essential data for the basin-wide spatio-temporal reconstruction of lacustrine and marginal lacustrine paleoenvironments. As successive reconstructions of paleolake history become increasingly refined, regional paleohydrologic and paleoclimatic reconstructions can become increasingly realistic. The scheme of paleolake analysis that is summarized in Figure 1 eventually culminates in applications as seemingly diverse as assessments of seismic hazards and flooding hazards.

In theory, closed-basin lakes can be regarded as very large storage gages, where hydrologic inputs and outputs--viz., basin precipitation minus basin evapotranspiration, lake precipitation, and lake evaporation--are summed algebraically over time. Theoretically, the resulting changes in lake surface area produce corresponding changes in lake stage that are inscribed on coastal landscapes as flights of shorelines, which ideally persist into the future as multiple-event staff gages or dip sticks. In practice, however, especially in geologically complex basins with large lakes, the dip-stick analogy tends to break down, and the stage changes that are represented geomorphically on basin slopes are often complicated by factors besides those that pertain to lacustrine water budgets. A variety of factors that can contribute to local stage changes or stillstands during the time of shoreline development are outlined in Table 1. Factors 1, 3, 5, and 6 (Table 1) contributed significantly to stage changes and stillstands that were recorded by Lake Bonneville shorelines. Furthermore, later isostatic rebound and faulting caused significant local deformation of the Lake Bonneville shoreline record following the time of shoreline development.

Fortunately, work with Lake Bonneville stratigraphy and Bonneville basin geology has progressed far enough to permit many of the local and basin-wide complicating factors to be diagnosed effectively.

Lake Bonneville Data

The chronometric and other stratigraphic information on which this synthesis is based is from the published literature and from unpublished data of the authors. Emphasis has been placed on a stratigraphic reconstruction of the littoral zone of Lake Bonneville, which in its spatio-temporal patterning is a record of lacustrine paleomorphometry. A chronostratigraphic framework is provided by a ladder of 50 selected radiocarbon dates, which for the most part are from stratigraphic contexts that are well known to the authors. Information pertaining to each of the dates is summarized in Table 2. Each of the dates is plotted in Figure 2, a time-altitude diagram of Lake Bonneville for the period from 32,000 to 10,000 yr B.P. Figure 2 includes downward adjustments of altitudes to compensate for the differential hydro-isostatic rebound of up to 74 m (243 ft) that occurred during and after unloading. Rebound-free adjusted altitudes, Za, have been estimated by the relationship

$$Za = Zr - [(Zr - 1200)/(Zb - 1200)][Zb - 1552],$$

where Zr is a particular rebounded altitude, Zb is the local altitude of the Bonneville Shoreline, and all values are in meters. The configuration of the Bonneville basin is such that an essentially linear relationship exists between rebound-free surface altitudes (Za) and the surface areas that occurred at intermediate and high levels of the lake (Figure 3). For values of Za above 1300 m, surface area, A, in square kilometers is a function of Za according to the regression equation

$$A = 137.2\ Za - 161400, \text{ with } r = 0.995 \text{ and } n = 4.$$

Summary of Lake Bonneville History

Important known stages and potentially important inferred stages in the history of Lake Bonneville are summarized in order of decreasing age in the following brief descriptions, which are keyed to Figure 2 by means of letters A through P.

A. Analyses of sediment cores from Great Salt Lake (Spencer et al., 1984) suggest that prior to about 32,000 yr B.P. the basin was occupied by an ephemeral shallow saline lake or playa. Oolitic sand and carbonate mud deposited in water less than 3 m (10 ft) deep and dated at about 31,660 yr B.P. (date 1) supports this interpretation.

B. Spencer et al. (1984) suggest that from about 32,000 to 25,500 yr B.P. the lake rose above shallow saline stages to altitudes close to 1294 m (4245 ft). Temporal control in the cores for this time interval is provided by two silicic volcanic ashes, the Carson Sink ash (29,000 yr B.P.) and the Wono ash (24,500 yr B.P.). However, the lake apparently rose somewhat faster than Spencer et al. (1984) suggest, and had transgressed to near 1292 m (4239 ft) by about 26,700 yr B.P. (date 2). By 26,000 yr B.P. (date 4) it was slightly above 1314 m (4311 ft).

C. Two fluctuations comprising the Stansbury oscillation, which had a maximum amplitude of about 45 m (150 ft), occurred between about 22,500 and 21,500 yr B.P. Disconformities in shallow lacustrine sediments at the Old River Bed record the fluctuations of the Stansbury oscillation (Oviatt, 1984; Oviatt, in preparation). Date 12 is on gastropods in sand and gravel of the Stansbury Shoreline at its type area on Stansbury Island (Currey et al., 1983), but as with several gastropod dates considered here, it appears too young when compared with wood dates. Stratigraphic evidence for the Stansbury Shoreline as a transgressive-phase feature is well documented (Currey et al., 1983; Oviatt, 1984), but its exact timing is still not well known.

D. An important oscillation or stillstand may have occurred between about 21,000 and 19,700 yr B.P. What seem to be significant accumulations of lagoonal sediments have been dated in this spatio-temporal interval at three localities.

E. A brief stillstand probably occurred about 19,500 yr B.P., when the main body of Lake Bonneville transgressed to the level of Sand Pass and spilled into the relatively dry Tule Valley subbasin. The event is clearly marked by a bed-load delta on the Tule Valley side of Sand Pass. Beach features that may be related to a Tule Valley stillstand occur about 11 m (36 ft) below the Provo Shoreline at many localities.

F. An important oscillation or stillstand may have occurred between about 19,000 and 18,500 yr B.P. Beach features belonging to this spatio-temporal interval seem to occur about 15 m (50 ft) above the Provo Shoreline at many localities, but none is better exposed than the massive transgressive lagoon/bar complex at the head of the Little Valley gravel pit. The term Pithead is applied here to his spatio-temporal interval. Contrary to the hypotheses of some workers, the basin-wide body of stratigraphic evidence that is directly associated with what has traditionally been regarded as the Provo stage of Lake Bonneville clearly dates from the regressive, and not the transgressive, phase of the last deep-lake cycle; a wealth of well-exposed stratigraphic evidence at dozens of localities is unequivocal on this point.

G. A very brief stillstand probably occurred shortly after 18,000 yr B.P., when the main body of Lake Bonneville transgressed to the threshold leading into the Cedar Valley subbasin. Evidence of scour by high-discharge flow is visible on the Cedar Valley side of the threshold. What may be the pre-inflow and post-Bonneville Flood shorelines of the indigenous Cedar Valley lake are also visible in the subbasin. Evidence of this event has not yet been identified with certainty outside the Cedar Valley area.

H. The average rate of Lake Bonneville transgression seems to have increased markedly after about 16,770 yr B.P. (date 26), during the culminating rise to external threshold control. The culminating rise may have been preceded by a stillstand or oscillation that is represented by the highest of the prominent beach features that are spatially intermediate between the Provo and Bonneville shorelines. The highest intermediate shoreline is typically about 27 m (90 ft) below the Bonneville Shoreline. Details of this apparent inflection (Figure 2) in the transgression of Lake Bonneville have yet to be resolved.

I. After about 16,000 years of intermittent expansion as a closed-basin hydrologic system, the culminating rise brought Lake Bonneville to external threshold control on its divide with the Snake River-Columbia River Basin, in southeastern Idaho. The controlling threshold was at 1552 m (5092 ft) above sea level, in a broad pass floored by Quaternary alluvial fans overlying Tertiary tuffaceous beds, near the present hamlet of Zenda. Initial attainment of threshold control at Zenda is not well dated, but prior and subsequent stratigraphic constraints suggest that it must have occurred about 16,400 yr B.P. Those constraints also suggest that threshold control persisted at least intermittently for about 500 years. At several localities, three discrete depositional components of the Bonneville Shoreline (Currey et al., 1983) seem to have been laid down during that interval, but it is not clear whether they resulted from three discrete subintervals of threshold control separated by two minor withdrawals or whether they were step-function responses to geomorphic process thresholds and isostatic subsidence (Table 1, factors 7 and 3b), with no interruption of threshold control. Morphostratigraphic details that are obscured by threshold-curtailed containment in the Bonneville basin may be represented by the high shorelines of other paleolakes in the eastern Great Basin.

J. A major drop in lake level from the Bonneville Shoreline to about 1508 m (4948 ft)--the Keg Mountain oscillation (Currey et al., 1983)--occurred between about 15,900 and 15,000 yr B.P. A basaltic ash in Bonneville sediments in the Sevier Desert provides temporal control near the low point of the oscillation (Oviatt, 1984). Three radiocarbon dates on gastropods and two on ostracodes from within or stratigraphically near the basaltic ash average about 15,300 yr B.P. (dates

29, 30, 31, 32, and 33), which here is considered a minimum limiting age for the ash. On Figure 2 the ash is plotted as slightly older than this average.

K. In completing the Keg Mountain oscillation, the lake rose to an overflowing stage at the Zenda threshold for a final time about 15,000 yr B.P. Date 34 provides a minimum estimate of the age of this short-lived, pre-Flood event.

L. The Bonneville Flood was unleashed into the drainage of the Snake and Columbia rivers about 15,000 years ago. In what was essentially a single event with some interesting but minor sub-events, the Zenda threshold was destroyed by about 108 m (355 ft) of downcutting and 3 km (2 mi) of headward erosion, probably in less than a year.

M. At the conclusion of the geologically instantaneous Flood, the Provo Shoreline stage of Lake Bonneville was under the control of a new external threshold at about 1444 m (4737 ft) above sea level, in the vicinity of what is now Red Rock Pass. Stratigraphic evidence for the Provo Shoreline as a regressive-phase feature is completely unambiguous in excellent exposures at many localities (e.g., Currey et al., 1983), but its timing is constrained only within moderately broad limits. Mollusc shells provide a minimum limiting age of about 13,900 yr B.P. (date 35) for late stages of Bear River delta development at the Provo Shoreline. A pattern of prograding and aggrading ramps that are terminated by small scarps (Figure 2) is conspicuous at many Provo Shoreline localities with extensive beach development. The authors now ascribe this morphostratigraphic signature of the Provo Shoreline to a Flood-triggered, persistently-active landslide of massive proportions that slowly raised the Red Rock Pass threshold, but which in turn was partially incised several times.

N. Following a climate-forced reversion to closed-basin hydrology about 14,000 yr B.P., the lake regressed steadily from the Provo Shoreline to at least as low as 1261 m (4137 ft) in approximately 2500 years. Although their exact stratigraphic contexts are not known, two wood dates (37 and 43) from offshore deposits seem to constrain the end of this drop to after 12,000 yr B.P.

O. After a maximum limiting date (38) of about 13,140 yr B.P. on organic carbon from fine-grained bottom sediments, up to 13 m (43 ft) of Glauber's salt (mirabilite) was deposited in the stratigraphic sequence that now underlies the deepest part of Great Salt Lake. The geochemistry of organic carbon in lacustrine sediments is such that date 38 probably constrains the onset of Glauber's salt precipitation to after 12,000 yr B.P. Spencer et al. (1984) suggest that the Glauber's salt began to be precipitated when the residual brines of Lake Bonneville were only a few meters deep.

P. After about 10,920 yr B.P. (date 47), a water body that was transitional between Lake Bonneville and Great Salt Lake expanded to the Gilbert Shoreline. However, Danger Cave, 15 m (50 ft) above the Gilbert Shoreline, has been continuously dry since the deposition of sheep dung about 11,450 yr B.P. (date 46). The Gilbert Shoreline may have resulted from two beach-forming events and seems to date from about 10,000 yr B.P. or slightly earlier (dates 48, 49, and 50).

Durations, Average Rates, and Probably Causes

From the foregoing summary, and from the data sets on which that summary is based, several tentative conclusions can be drawn regarding durations, average rates, and probable causes of Lake Bonneville expansions, stillstands, and contractions (Table 3). Although data are still sparse and knowledge is very incomplete, the tentative conclusions developed in this study can be advanced as a coherent set of hypotheses to help guide additional Lake Bonneville research in the immediate future. The apparent regularity of the basin-wide hypsimetric curve at altitudes above 1300 m (Figure 3) simplifies the interpretation of Lake Bonneville history; if real, this regularity permits direct comparisons of stage-change rates at various levels above that altitude.

Several generalizations can be made regarding the tentative conclusions presented in Table 3 (and Figure 2). The apparent lack of significant oscillations during the post-Provo contraction may well be real, a logical consequence of persistent water budget deficits caused by postglacial (postlacustral) climatic forcing. The apparent lack of significant oscillations during the pre-Stansbury expansion may well be an artifact of sparse data, and the actual pattern may more nearly resemble the succession of significant oscillations that occurred subsequently. The average rates of contraction and expansion, averaged together because of limited spatio-temporal resolution of the stratigraphic record, that seem to have been typical of Lake Bonneville oscillations were on the order of 10 to 15 sq km (4 to 6 sq mi) per year. There is no reason to believe that oscillations with average rates on that order could not have occurred during the pre-Stansbury expansion. A better understanding of the long history of the pre-Stansbury expansion is highly relevant to a better understanding of potential post-Great Salt expansion stages of the next deep-lake cycle.

The maximum potential extent of Lake Bonneville can be estimated, but will never be known with certainty because of intervention by the Zenda threshold. On the basis of the reconstruction presented here, it seems unlikely that even with complete containment the lake would have greatly exceeded a maximum stage of about 1570 m (5150 ft) or a maximum surface area of about 54,000 sq km (20,850 sq mi).

Stillstands in the literal sense were very rare, even when lake stages were threshold controlled. Despite control by the Zenda threshold, continued hydro-isostatic subsidence seems to have exerted a significant influence on the configuration of the developing Bonneville Shoreline in the basin interior prior to the Keg Mountain oscillation; isostatic rebound during the Keg Mountain oscillation, when about one-quarter of the maximum Lake Bonneville water load was removed and then largely but not completely replaced, seems to have exerted a significant influence on the configuration of the specific part of the Bonneville

Shoreline that was occupied at the start of the Bonneville Flood. At Red Rock Pass, massive post-Flood landsliding from the east flank of the Malad Range seems to have interacted with downcutting by Lake Bonneville overflow to strongly influence the morphostratigraphic development of the Provo Shoreline; cumulative rise of the Red Rock Pass threshold due to landsliding slightly exceeded cumulative downcutting, and even seems to have kept pace with basin-interior rebound.

The overall time-stage hydrograph of the Bonneville deep-lake cycle is clearly asymmetric (Figure 2), with a net expansion rate of about 3.3 sq km (1.3 sq mi) per year during the 16,000-year transgressive phase, and a net contraction rate of at least 13.2 sq km (5.2 sq mi) per year during the regressive phase of less than 4,000 years. In other words, the net rate of regressive-phase contraction was at least four times greater than the net rate of transgressive-phase expansion. This asymmetry of the Bonneville basin hydrograph during the last deep-lake cycle is very similar to the asymmetry of the last major glaciation as recorded in deep-sea sediments during oxygen isotope stage 2 of the world oceanic chronology. In particular, the rapid decline of Lake Bonneville about 13,000 yr B.P. seems to have been exactly synchronous with the termination of stage 2, when the isotopic signal in deep-sea sediments switched rapidly from glacial to postglacial.

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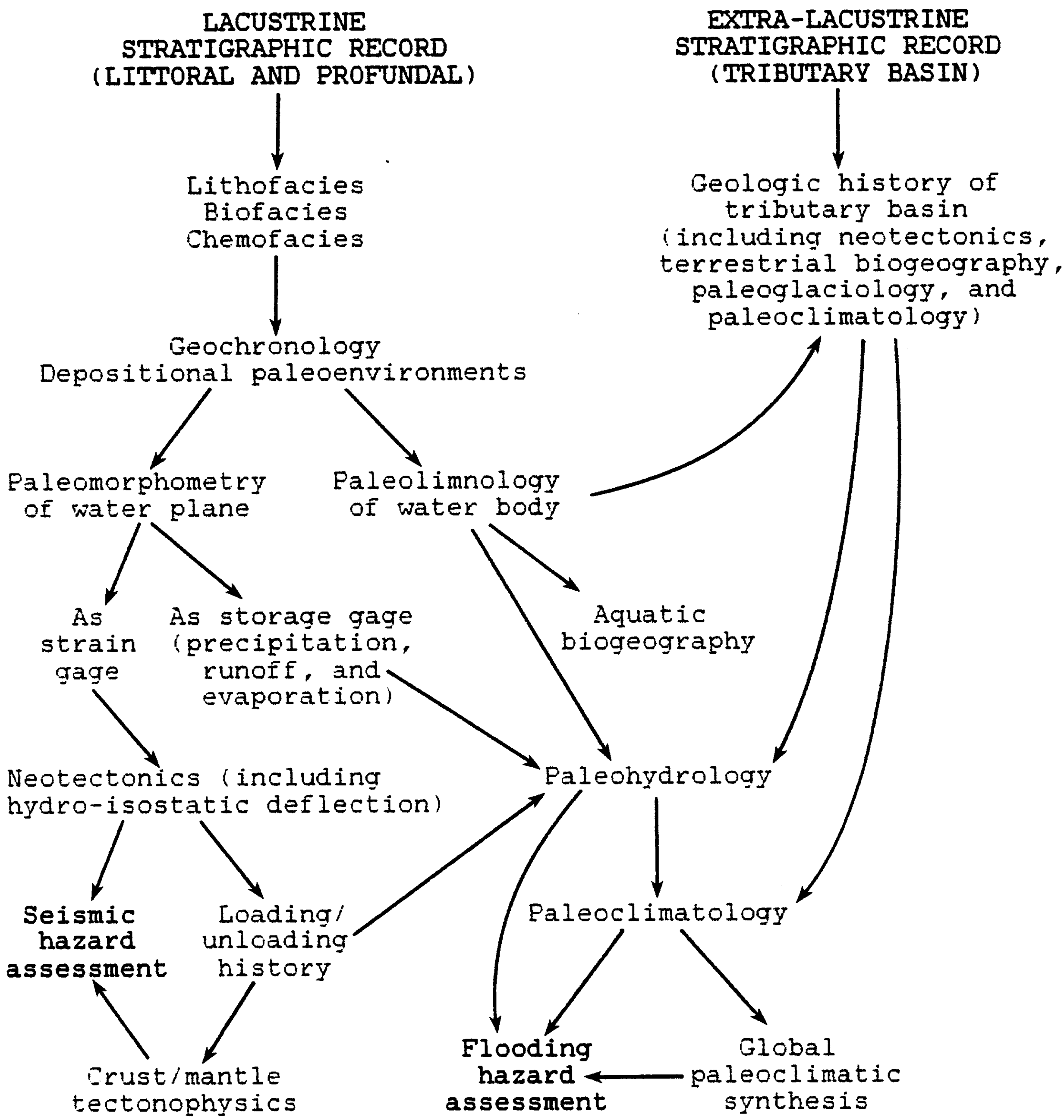
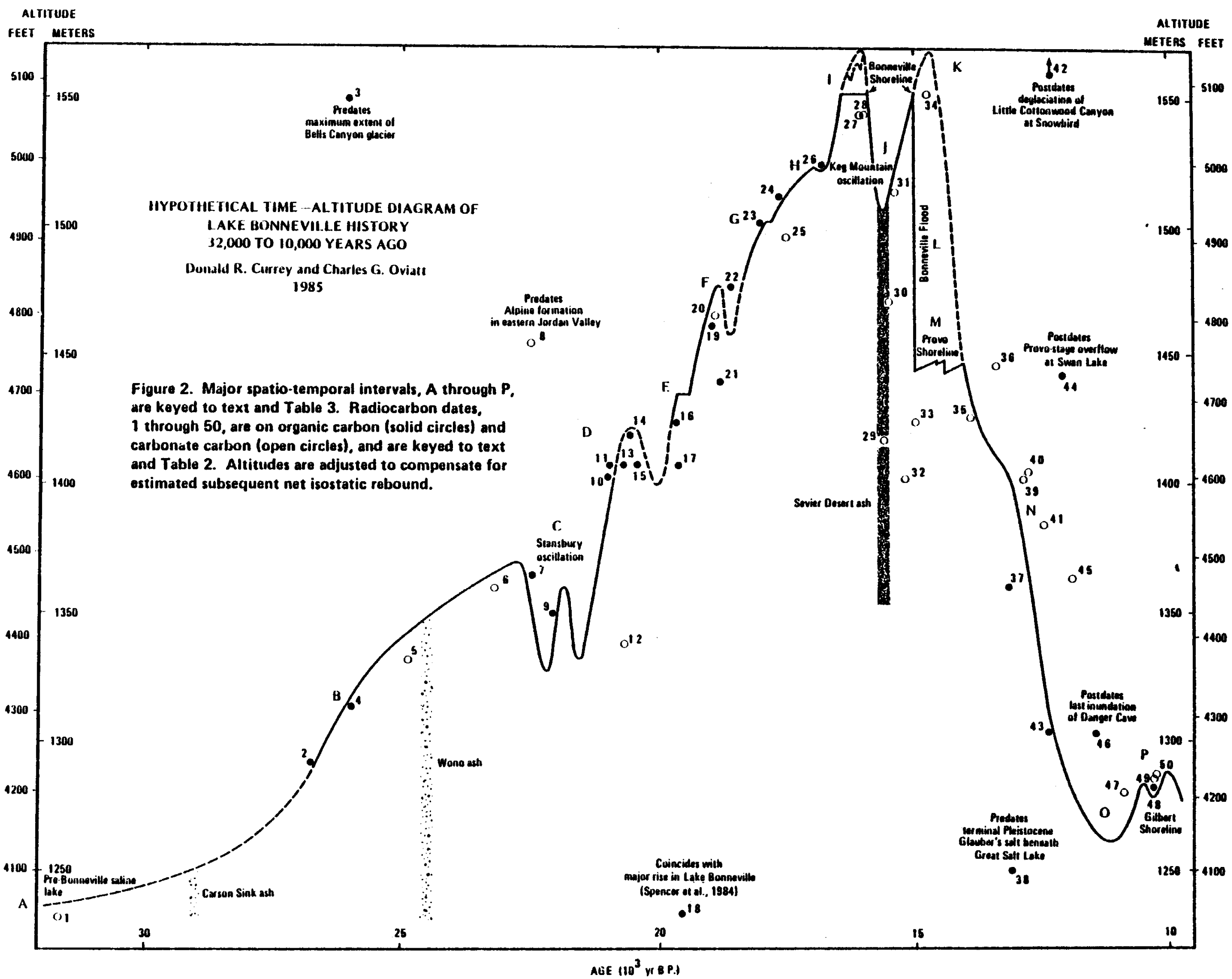


Figure 1. Typical scheme for basin-wide reconstruction and interpretation of late Pleistocene and Holocene closed-basin lakes, with applications to hazard assessment, including possible flooding by unusually high lake stages.



HYPOTHETICAL TIME—ALTITUDE DIAGRAM OF
LAKE BONNEVILLE HISTORY
32,000 TO 10,000 YEARS AGO

Donald R. Currey and Charles G. Oviatt
1985

Figure 2. Major spatio-temporal intervals, A through P, are keyed to text and Table 3. Radiocarbon dates, 1 through 50, are on organic carbon (solid circles) and carbonate carbon (open circles), and are keyed to text and Table 2. Altitudes are adjusted to compensate for estimated subsequent net isostatic rebound.

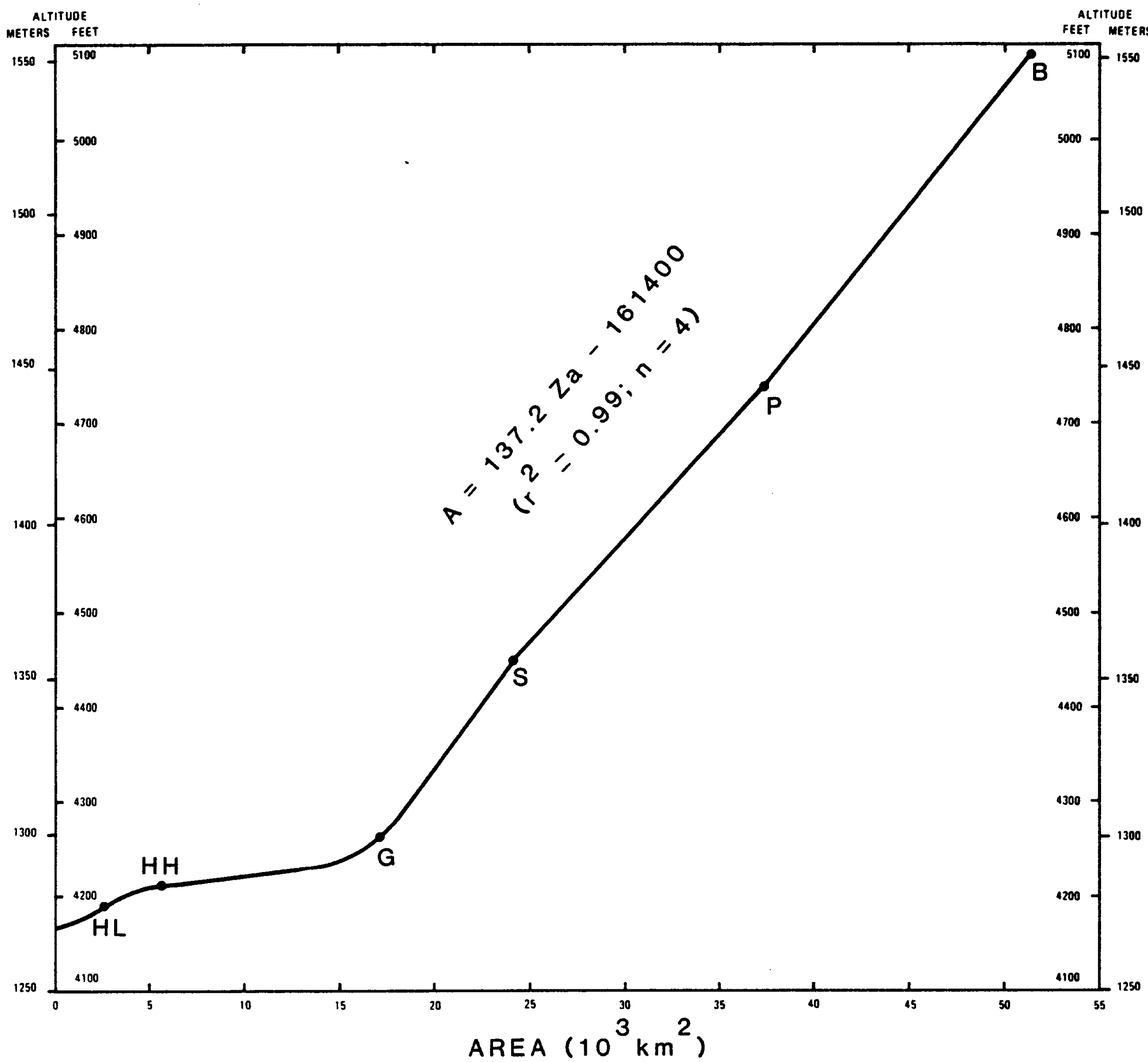


Figure 3. Hypsimetric diagram showing stage areas of Great Salt Lake and Lake Bonneville. Stages symbolized are the historic low (HL), historic high (HH), Gilbert Shoreline (G), Stansbury Shoreline (S), Provo Shoreline (P), and Bonneville Shoreline (B). Stage altitudes above 1295 m (4250 ft) are adjusted to compensate for estimated subsequent net isostatic rebound.

TABLE 1. Factors That Can Contribute to Local Stage Changes or Stillstands in a Large Closed-Basin Lake.

1. **Climatic forcing** (causing expansion, stillstand, or contraction of lake area)
   a. Hydroclimatic inputs increase, remain constant, or decrease
   b. Hydroclimatic outputs decrease, remain constant, or increase

2. **Basin configuration** (governing stage change as a function of lake area change)
   a. Flatter-than-average reach of basin area-altitude curve
   b. Steeper-than-average reach of basin area-altitude curve

3. **Basin deformation** (causing local stage change independent of lake area change)
   a. Fault-block tectonism
   b. Hydro-isostatic deflection (especially in basin interior)
   c. Litho-isostatic deflection (in depocenter area)

4. **Basin area change**
   a. Enlargement of runoff-producing area by stream capture
   b. Reduction of runoff-producing area by stream capture

5. **Basin hydrographic threshold**
   a. External threshold curtails potential expansion
   b. Internal threshold curtails potential expansion in spilling subbasin and augments expansion in receiving subbasin

6. **Basin hydrographic threshold change**
   a. Downcutting tends to lower threshold
   b. Landsliding tends to raise threshold
   c. Volcanic eruptions tend to raise threshold
   d. Faulting can raise or lower threshold
   e. Isostatic deflection can raise or lower threshold

7. **Coastal geomorphic process threshold** (causing step-function development of depositional form, because of abrupt transition in regime variables)
   a. Wave dynamics
   b. Sediment supply
   c. Shore-zone topography

8. **Simultaneous influence of multiple climatic factors and multiple geologic factors**

TABLE 2. Radiocarbon Dates Used in Diagram of Lake Bonneville History, 32,000 to 10,000 Years Ago

Fig. 2 No.	Lab No.	Age ± s.d. (y)	Material	Locality	Present altitude (m)	Estimated original altitude (m)	Stratigraphic context	Reference(s)
1	Beta-11489	31,660 ± 740	Oolitic sand with carbonate mud matrix	Southern Pacific causeway borehole in Great Salt Lake	1232	1232	Shallow saline lake predating last deep-lake cycle	This paper
2	I-4409	26,700 ± 900	Dark gray calcareous sediments	Crescent Spring	1292	1292	Spring-fed marsh predating last deep-lake cycle	Mehringer, 1977
3	GX-4737	26,080 ± 1150	Soil organic carbon	Majestic Canyon Estates near mouth of Bells Canyon	1550	1550	Predates maximum extent of Bells Canyon glacier, which predates last deep-lake cycle	Madsen and Currey, 1979; Currey, 1980
4	W-4893	26,000 ± 600	Wood	Bullion Street pit, Murray	1314	1314	Transgressive delta	Scott et al., 1983
5	Beta-8343	24,870 ± 410	Aragonitic marl, < 10 micron fraction	Stansbury Gulch, S end Stansbury Is.	1356	1333	Transgressive off-shore zone	This paper
6	Beta-5038	23,190 ± 1360	*Sphaerium* shells	Old River Bed near Shutoff	1378	1359	Transgressive near-shore zone	Oviatt, 1984; Currey et al., 1983
7	W-4898	22,500 ± 300	*Picea* wood	Moffat Bros. pit, Kearns	1379	1365	Transgressive gravel-bar complex	Scott et al., 1983
8	Beta-8344	22,500 ± 370	Soil carbonate carbon, < 10 micron fraction	Dimple Dell Geosol at type locality	1455	1455	Predates local transgression of last deep-lake cycle	Currey et al., 1984
9	I-698	22,000 ± 600	Wood	Logan water well	1357	1349	Transgressive delta (?)	Trautman and Willis, 1966
10	W-4897	20,900 ± 250	*Abies* or *Juniperus* wood	Pit N of Jordan Narrows	1417	1405	Transgressive lagoon/bar complex	Scott et al., 1983
11	L-775N	20,800 ± 300	Wood	Little Valley	1442	1408	Transgressive shore zone	Morrison, 1965
12	Beta-5566	20,710 ± 310	*Amnicola* shells	Stansbury Gulch, S end Stansbury Is.	1362	1338	Nearshore zone traceable to type Stansbury Shoreline	Currey et al., 1983
13	W-876	20,600 ± 500	Wood	Little Valley	1442	1408	Transgressive shore zone	Rubin and Alexander, 1960
14	SI-4124	20,500 ± 200	Wood	CPC pit, N Salt Lake	1438	1417	Transgressive lagoon/bar complex	R. Stuckenrath, pers. com. 1979
15	W-941	20,300 ± 300	Wood	Little Valley	1442	1408	Transgressive shore zone	Ives et al., 1964; Morrison, 1965
16	W-4421	19,700 ± 200	*Picea* wood	CPC pit, N Salt Lake	1445	1424	Transgressive lagoon/bar complex	Scott et al., 1983
17	W-4445	19,580 ± 280	*Abies* or *Juniperus* wood	Little Valley	1442	1408	Transgressive lagoon/bar complex	Scott et al., 1982; Scott et al., 1983
18	W-4779	19,540 ± 150	Wood	Core J, S of Promontory Point	1269	1234	Transgression from unit IIId to unit IIIc	Spencer et al., 1984
19	SI-4041C	18,980 ± 165	Wood	Monroc-Stauffer pit, N Salt Lake	1487	1462	Transgressive lagoon/bar complex	R. Stuckenrath, pers. com., 1978
20	W-982	18,980 ± 500	*Stagnicola* shells	Ramsbottom Ranch, Idaho	1478	1466	Transgressive near-shore zone (?)	Bright, 1963; Ives et al., 1964
21	W-4695	18,800 ± 180	Wood	CPC pit, N Salt Lake	1463	1440	Transgressive lagoon/bar complex	Scott et al., 1983
22	W-4693	18,600 ± 150	*Picea* wood	Pit at Point of the Mountain	1494	1477	Transgressive shore zone	Scott et al., 1983
23	W-4687	18,000 ± 150	Wood	Holladay Gun Club	1524	1502	Transgressive shore zone	Scott et al., 1983
24	W-4451	17,580 ± 170	Charcoal	Holladay Gun Club	1535	1512	Transgressive shore zone	Scott et al., 1983; Currey and James, 1982
25	L-711C	17,500 ± 400	Gastropod shells	Leamington amphitheatre	1507	1496	Transgressive near-shore zone	Broecker and Kaufman, 1965; Oviatt, 1984

26	W-4896	16,770 ± 200	Wood	MacNeish pit, N Salt Lake	1553	1523	Transgressive shore zone	Scott et al., 1983
27	L-363G	16,100 ± 350	Tufa	N end of Oquirrh Mts.	1585	1544	Near Bonneville Shoreline	Broecker and Kulp, 1957
28	L-4838B	16,050 ± 300	Tufa	N end of Oquirrh Mts.	1585	1544	Near Bonneville Shoreline	Olson and Broecker, 1961
29	L-711B	15,600 ± 400	Ostracode shells	SE bank of Sevier River W of Fool Creek Reservoir No. 2	1427	1417	0.13 to 0.26 m below Sevier Desert ash, in white marl	Broecker and Kaufman, 1965
30	L-774H	15,500 ± 500	Gastropod shells	Marl pit SE of Fool Creek Reservoir No. 2	1480	1471	0.08 to 0.30 m above Sevier Desert ash, in white marl	Broecker and Kaufman, 1965
31	L-774N	15,400 ± 300	Gastropod shells	Leamington amphitheatre	1525	1514	Brackets Sevier Desert ash, in white marl	Broecker and 1965; Oviatt, 1984
32	Beta-10389	15,220 ± 140	Ostracode shells	DMAD Dam	1420	1402	At base of Sevier Desert ash, in white marl	This paper
33	L-774F	15,000 ± 400	Gastropod shells	SE bank of Sevier River, W of Fool Creek Reservoir No. 2	1433	1424	Within Sevier Desert ash in white marl	Broecker and Kaufman, 1965
34	SI-4227C	14,730 ± 100	Tufa, innermost 18%, C13/C12 adjusted	Stockton Bar	1579	1552	Youngest occupation of Bonneville Shoreline	R. Stuckenrath, pers. com., 1979, Currey et al., 1983
35	W-899	13,900 ± 400	Mollusc shells	W of Preston, Idaho	1433	1426	Bear River paleo-delta graded to Provo Shoreline	Rubin and Berthold, 1961; Bright, 1963
36	W-491	13,380 ± 400	Tufa	N end of Stansbury Mts.	1494	1448	Provo Shoreline	Rubin and Alexander, 1958
37	I-697	13,150 ± 270	Wood	Logan water well	1367	1360	Post-Provo regressive delta (?)	Trautman and Willis, 1956
38	Beta-10246	13,140 ± 650	Organic carbon in pelagic ooze	Southern Pacific causeway borehole in Great Salt Lake	1261	1250	0.6 m below base of Glauber's salt	This paper
39	W-2000	12,860 ± 400	Gastropod shells	Black Rock Canyon, N end Oquirrh Mts.	1424	1402	Regressive nearshore zone (?)	Smith et al., 1968; Marsters et al., 1969
40	W-943	12,780 ± 350	Gastropod shells	Little Valley	1440	1404	Regressive nearshore zone (?)	Ives et al., 1964; Morrison, 1965
41	Beta-8348	12,490 ± 130	<u>Anodonta</u> shells	Sunstone Knoll, Sevier Desert	1396	1384	Regressive nearshore Zone	Isgreen et al., 1984
42	GX-3481	12,300 ± 330	Wood	Snowbird bog	2470	2470	Basal peat overlying late Pleistocene ground moraine	Madsen and Currey, 1979; Currey, 1980
43	W-1824	12,290 ± 350	Wood Chips	Roy	1313	1304	Regressive (?) upward coarsening sequence	Ives et al., 1967
44	W-1338	12,090 ± 300	Plant fragments	Swan Lake, Idaho	1442	1442	Postdates Provo-stage overflow at Swan Lake	Bright, 1966
45	L-774Q	11,900 ± 300	Ostracode shells	Old River Bed at Shutoff	1382	1363	Regressive nearshore zone	Broecker and Kaufman, 1965; Currey et al., 1983
46	C-609	11,453 ± 600	<u>Ovis</u> dung	Danger Cave	1314	1303	Postdates last inundation of Danger Cave	Libby, 1955; Jennings, 1957
47	W-4395	10,920 ± 150	Gastropod shells	Public Shooting Grounds	1289	1280	Marsh predating rise to Gilbert Shoreline	Miller, 1980
48	I-696	10,300 ± 275	Wood	Willard canal	1290	1283	Overlain by sand of Gilbert transgression	Trautmen and Willis, 1966
49	GX-6949	10,300 ± 310	Gastropod shells	Magna spit	1294	1286	Marsh interbedded in shallow lagoon at Gilbert Shoreline	Currey et al., 1983
50	GX-6614	10,285 ± 265	Gastropod shells	Magna spit	1294	1286	Marsh interbedded in shallow lagoon at Gilbert Shoreline	Currey, 1980; Currey et al., 1983

TABLE 3. Durations (D), Average Rates (R), and Probable Causes (C) of Lake Bonneville Expansions, Oscillations, Stillstands, and Contractions. (Letters B through P are keyed to Figure 2.)

B. **Pre-Stansbury expansion** R: + 3 sq km per yr	D: 5,000+ yr C: climate
C. **Stansbury oscillations 1 & 2** R: - & + 14 sq km per yr (?)	D: 1,400 yr C: climate
D. **Post-Stansbury oscillation (?)** R: (?)	D: 1,000 yr (?) C: climate (?)
E. **Tule Valley stillstand** R: negligible in main body	D: 200 yr (?) C: internal threshold with significant downcutting
F. **Pithead oscillation** R: (?)	D: 500 yr (?) C: climate
G. **Cedar Valley stillstand** R: negligible in main body	D: 100 yr (?) C: internal threshold with slight downcutting
H. **Highest intermediate oscill'n** R: (?)	D: 300 yr (?) C: climate
I. **Bonneville "stillstand"** R: + 1 sq km per yr	D: 500 yr C: external threshold with basin-interior subsidence & insignificant downcutting
J. **Keg Mountain oscillation** R: - & + 12 sq km per yr	D: 900 yr C: climate
L. **Bonneville Flood** R: - 14,000 sq km per yr	D: 1 yr C: external threshold with catastrophic downcutting
M. **Provo "stillstand"** R: minor	D: 1,000 yr (?) C: external threshold with landsliding & downcutting
N. **Post-Provo contraction** R: - 13 sq km per yr	D: 2,500+ yr C: climate
P. **Gilbert oscillation(s)** R: + & - (?)	D: 1,000 yr (?) C: climate

Late Holocene Lake-Level Fluctuations of the Great Salt Lake (Utah) as Deduced from Oxygen-Isotope and Carbonate Contents of Cored Sediments

Judith A. McKenzie*
Gregor P. Eberli

Geological Institute
Swiss Federal Institute of Technology
8092 Zurich, Switzerland

*Present Address: Department of Geology, University of Florida, Gainesville, FL 32611

INTRODUCTION

The Great Salt Lake, Utah is situated topographically at the bottom of a closed basin. Lacking an outlet, the water level is controlled by a balance between an inflow of freshwater by rivers, groundwater, and direct precipitation versus an evaporative outflow. Historic records kept since 1851 A.D. show that the lake level has fluctuated as much as 6.4 m (21 ft.) around a mean elevation of 1280.8 m (4202 ft.). The historic high in 1873 AD was at 1283.8 m (4212 ft.) while the lowest recorded level was at 1277.4 m (4191 ft.) in 1963 AD (Currey et al., 1984). Since this low period, the lake has risen gradually with a sudden increase in the ascent rate beginning in 1983 and is expected to peak at 1283.7 m (4211.5 ft.) in 1985 (Arnow, this volume).

The pre-historic record obtained from studies of ancient topographic features around the lake and lacustrine sediments indicates that this dynamic situation is the norm for the Great Salt Lake. Since the latest Pleistocene the lake level has reached several, recognized high stands. During the most notable high stand, the Lake Bonneville Stage at 1551.5 m (5090 ft.) about 16,000-14500 years B.P., the waters reached the threshold near Red Rock Pass, Idaho and spilled into the Snake River converting the lake temporarily into an open system (Currey et al., 1984). Thus, from both historic and pre-historic records the evidence is overwhelming that the lake level can and does fluctuate widely. Predicting how much and how fast it will fluctuate remains an important area of investigation.

The purpose for this study of sediment cores taken from the Great Salt Lake was to evaluate the oxygen-isotope stratigraphy of the authigenic carbonates in order to obtain a record of lake-level fluctuations. The oxygen-isotope composition of natural waters can vary greatly as a result of isotope fractionation during evaporative/precipitaton processes in the hydrologic cycle (Epstein, 1959). Meteoric or continental waters derived from precipitation are, in general, depleted in the heavier isotope (oxygen-18), while during evaporation, the oxygen-18 concentration in a water body tends to increase. We proposed that the oxygen-isotope composition of the Great Salt Lake waters should vary as a function of the ratio of fresh-water input versus evaporative output, i.e. an increase in the rate of fresh-water input would result in an oxygen-18 decrease in the lake waters while conversely an increase in the rate of evaporative output would enrich the lake waters in oxygen-18. These oxygen-18 variations would be incorporated in the authigenic carbonates at the time of their precipitation and with deposition an indirect record of lake-level fluctuations would be stored in the lacustrine sediments. Also, with a significant decrease in salinity, a change over from aragonite to calcite precipitation should occur. This paper reports the results of our investigation of oxygen-isotope and carbonate stratigraphies in late Holocene sediments from the Great Salt Lake.

METHODOLOGY

In 1979 sediment cores were secured from the south arm of the Great Salt Lake for a multi-disciplinary study of lake-level history. Stratigraphic correlations of the various types of data (sedimentology, stable isotope, organic geochemistry, pollen, ostracode assemblages etc.) were used to interpret latest Pleistocene/Holocene lake elevation (Spencer et al., 1984; McKenzie et al., 1982; McKenzie and Eberli, 1984). For this study of late Holocene lake levels, we chose cores from Site I, which lies west of Antelope Island at about 41°N, 112 20'W (Figure 1). Sedimentary correlations showed that cores from this site contained an expanded sequence above the highest tephra layer, which is correlated with the Mazama ash dated at 6,900 years B.P. The sediments comprise laminated, aragonitic alterations of sorted pellets, ripple laminae, unsorted parallel flat laminae and disrupted laminae (Spencer et al., 1984). It was anticipated that greater stratigraphic resolution could be obtained from such a laminated, expanded section.

A piston core (GSL-79-05P2) was sampled every 5 cm over 0.5 cm intervals in order to achieve an unbiased delineation of isotopic variations with depth in the core. After the analyses of these initial samples was successfully completed, the upper 0.45 m of the piston core were resampled at approximately 2 cm intervals to obtain a more detailed isotope record for the latest Holocene sediments. The upper 0.44 m of a gravity core (GSL-5G2) with less coring disturbance than the piston core was likewise densely sampled. With this dense sampling of two cores, we hoped to obtain an isotope record with enough continuity and detail that it could be correlated with the historic record of lake-level fluctuations.

Each sample was treated first with a 6% sodium hypochlorite solution to remove any organic matter. Afterwards, the washed and dried samples were analyzed by X-ray diffraction techniques. The total weight percentage of carbonate in the treated samples was also measured using a combustion-titration apparatus, which in combination with the X-ray data enabled the calculation of the percentage of aragonite, calcite, dolomite and quartz in the samples. Stable isotope analyses of CO2 gas released from the samples during reaction with phosphoric acid at 25°C under vacuum were made with a triple-collector mass spectrometer, VG Micromass 903. The oxygen-isotope ratio of the bulk carbonate in the sample is reported as the per mil (o/oo) deviation from the international PDB isotopic standard. The pertinent data are shown in Figures 2 and 3 and will be listed in a future, comprehensive publication on the results of our entire study.

RESULTS AND DISCUSSIONS

Source of Oxygen-18 Variations

The first test of the applicability of oxygen-isotope stratigraphy in the study of lake-level fluctuations was to determine the sedimentary source of the oxygen-isotope variations. Our data for the late Holocene, as shown in Figure 2 for the gravity core, consistently showed that the weight percentage of the detrital carbonates, calcite and dolomite, remained constant while the weight percentage of aragonite and total carbonate varied in a parallel manner, exclusive of periods of very low salinities when calcite is the stable authigenic precipitate. Further, the oxygen-18 composition tended to vary analagously, i.e. increasing percentage aragonite (or percentage carbonate) are associated with increasing oxygen-18 values, while decreasing percentage aragonite (or percentage carbonate) correspond to decreasing oxygen-18 values. If we assume that the oxygen-18 composition of the authigenic aragonite reflects the isotope ratio of the lake water at the time of precipitation, we can then correlate periods of increasing oxygen-18 and percentage aragonite (or percentage carbonate) with periods of increased evaporative output and periods of decreasing oxygen-18 and percentage aragonite (or percentage carbonate) with periods of increasing fresh-water input.

Determination of Stratigraphic Ages

The second test of our proposed relationship between oxygen-isotope stratigraphy and lake-level fluctuations was to compare the isotopic record with the historic record. To date, the age control in the cores has been based on very limited data. The presence of the Mazama tephra layer at 4.38 m in the 5.01 m piston core places a maximum age of 6900 years B.P. near the base of the core. The minimum age (1978 A.D.) was set at the top of the gravity core. A third, very approximate age of 5500 years B.P. is set at 1.825 m, where the last appearance of authigenic dolomite occurs in the sediments. A mid-Holocene playa stage (7000 to 5500 years B.P.) has been recognized for the region containing the Great Salt Lake Basin (Curry and James, 1982). We correlate this playa stage with the frequent appearance of authigenic dolomite in the sediments. Above 1.825 m the authigenic carbonate is dominantly aragonite with the exception of a brief interval around 0.95 m where calcite becomes the dominant authigenic carbonate. We associate this latter mineralogic change with the mid-Neolacustral episode which occurred between 3,500 and 2,200 years B.P. and is recognized by an incursion of lacustrine facies in a sequence of salt marsh sediments (Mehringer, 1977; Currey and James, 1982).

Placing approximate ages of 5,500 and 2,200 years B.P., respectively, at these two points of mineralogic change (1.825 and 0.95 m), an average sedimentation rate of 0.26 mm/yr can be calculated. The magnitude of this rate is common for saline lacustrine systems (G.I. Smith, personal communication), and we have used it to extrapolate our stratigraphic ages between 5,500 and 1,000 years B.P. (Figure 3). Since compaction of the younger sediments will not be as extensive as in the older sediments, we distribute the last 1,000 years between 0.65 m and the top of the core taking into account our correlation of the isotope stratigraphy with the historic record.

In Figure 2, the historic lake-level record is correlated with the oxygen-isotope record contained in the upper 45 cm of cored sediment. Although this correlation remains only very approximate due to the lack of absolute dating of the cores, a strong similarity between the oscillations in the isotopic and historic curves apparently exists. In particular, the shape of the isotopic curve for the upper most 22 cm of the gravity core shows a remarkable resemblance to that from the historic record. The most negative $\delta^{18}O$ value at 20 cm could correspond to the highest recorded stand in 1873 A.D., while the most positive $\delta^{18}O$ value at 5 cm could record the historic low stand in 1963 A.D. It is undoubtedly significant that the intervening isotope data between 20 and 5 cm indicate two low and two high stands between the 1873 and 1963 A.D. extremes. These isotopic fluctuations could be a manifestation of the 1905 and 1945 A.D. lower stands and 1925 and 1952 A.D. higher stands, as seen in the historic lake-level record. Although not as complete or distinctive, the isotope data for the piston core shows comparable trends. Together, the data from lake cores strongly suggest that the oxygen-isotope signal recorded in the authigenic carbonate does reflect changes in lake elevation.

Rate of Oxygen-18 Change vs. Lake-Level Fluctuations

If we assume that the variations in the oxygen-18 curve correspond to fluctuations in the historic lake-level curve, we can calculate the rate of oxygen-18 decrease or increase ($\Delta \delta^{18}O$) with the respective rises or falls in the lake level (Δ h). An absolute rate of change in oxygen-18 versus change in elevation ($|\Delta\delta^{18}O/\Delta h|$) can, thus, be obtained from the available data. The data are compiled in Table 1. In Figure 2, the maximum and minimum $\delta^{18}O$ values are correlated with the dates of the respective low and high stands in the historic record. The $\delta^{18}O$ values, of course, do not represent annual values but are a mixture of values for an indeterminant number of years, a result of sampling over 0.5 cm intervals. Using a total of 11 data sets from both the piston and gravity cores, it was possible to determine an average rate of change ($\Delta\delta^{18}O/\Delta h$) of 0.35 o/oo/m (0.11 o/oo/ft). This rate suggests that for each meter rise or fall in lake elevation the $\delta^{18}O$ value of the authigenic carbonate will change by 0.35 o/oo. Further, we calculated an average $\delta^{18}O$ value of -6.06 o/oo for the historic record from all of the available data points (27) which were correlated as being post 1851 A.D. This $\delta^{18}O$ value is considered to be representative of the modern mean elevation of 1280.8 m (4202 ft). Inasmuch as the lake level fluctuates around this mean elevation, a higher stand due to increased fresh-water input will result in a $\delta^{18}O$ value more negative than -6.06 o/oo, and, on the contrary, the $\delta^{18}O$ value will be more positive than -6.06 o/oo when evaporative output dominates.

Boundaries can be placed on the range within which the $\delta^{18}O$ value of the authigenic sediment can vary. At the Pleistocene/Holocene boundary, a rapid lowering of the lake from an extremely high elevation with fresh water conditions to a hypersaline playa system was recognized in sediment cores (Spencer et al., 1984). With the drawdown, there was a progressive increase in salinity and the authigenic carbonate changed from calcite to Mg-calcite to aragonite. There was also a linear increase in the $\delta^{18}O$ composition of the sediment from -8.1 to -1.7 o/oo (Spencer et al., 1984). If the lake waters become fresh enough to precipitate calcite, perhaps, the oxygen-18 composition of the water had reached a value that reflects predominantly a fresh-water input with relatively little evaporative fractionation. Therefore, a $\delta^{18}O$ value of about -8.0 o/oo is the minimum expected for the late Holocene sediments in the Great Salt Lake and will only be found when the lake is fresh enough to precipitate calcite as the authigenic carbonate. Using our rate of change factor of 0.35 o/oo/m, a $\delta^{18}O$ value of -8 o/oo will occur once the lake reaches a level of about 1286.5 m (4221 ft) or about 5.7 m above the mean modern elevation (1280.8 m).

The maximum $\delta^{18}O$ value which could be expected is about -2 o/oo, as indicated by the Pleistocene/Holocene isotope record (Spencer et al., 1984). The bottom of the Great Salt Lake lies at about 1271.0 m (4170 ft). Using the rate of change factor, a drop of 8.6 m below the modern mean elevation to 1272.2 m (4173 ft) would increase the $\delta^{18}O$ value of the authigenic aragonite to -3 o/oo. Assuming that these extrapolations are correct, the $\delta^{18}O$ values of the carbonate precipitating from the late Holocene Great Salt Lake can vary due to changes in the ratio of input versus output between approximately -8 o/oo and -3 o/oo, representing respectively a high stand of 1286.5 m (4221 ft) and a low stand of 1272.2 m (4173 ft). In Figure 3, we have superimposed these isotope lake-levels on the oxygen-isotope curve, i.e. -8 o/oo equates to 1286.5 m, -3 o/oo equates to 1272.2 m, and -6 o/oo represents the modern mean elevation 1280.8 m. The lake can rise higher than 1286.5 m but the oxygen-isotope composition will no longer be significantly altered, as the $\delta^{18}O$ value reflects basically the regional meteoric value. If the lake drops below 1272.2 m, the environment most likely becomes that of an ephemeral playa lake, and the $\delta^{18}O$ value of the authigenic carbonates no longer reflects the lake's elevation.

THE LATE HOLOCENE ISOTOPE RECORD

Neolacustrine Episode

Holocene paleoenvironments of the northeastern Great Basin, which includes the modern Great Salt Lake, have been extensively analysed and correlated utilizing litho- and morphostratigraphies, pollen and macrofossil stratigraphies and radiocarbon dating (Summarized in Currey and James, 1982). These correlations have lead to a consistent interpretation of the paleoclimaic changes in the region during the late Holocene, the period under consideration in this study. After a long dry period (7500-5000 years B.P.) when the Great Salt Lake experienced near desiccation, the climate of the region apparently became more amenable with the onset of a neopluvial episode, which has been subdivided into three glacial-lacustral-pluvial maxima and two minima (Currey and James, 1982). The early, mid- and late neopluvial stages or maxima are correlated with periods of higher lake levels and cooling coinciding with increased westerly flow and winter precipitation beginning at 5000, 3500 and 600 years B.P., respectively. The minima are associated with warmer periods of strengthened monsoon circulation, increased summer precipitation and wide-spread mesic conditions, beginning at about 4500 and 1500 years B.P. (Currey, 1976).

The late Holocene oxygen-isotope stratigraphy in sediment cores from the Great Salt Lake (Figure 3) is basically in agreement with the climatic record compiled from the numerous studies, as summarized by Currey and James (1982). Combining both the isotope and carbonate data for the lacustrine sediment record and using the approximated sedimentation rate and the calculated rate of oxygen-18 variation with changes in lake level, the elevation of the Great Salt Lake for the past 5500 years can be divided into three periods of higher stands with two intervening periods of lower stands. The estimated stratigraphic ages placed on these intervals are consistent with the general climatic record for the region.

Early Neolacustrine Maxima-5 (5500-3600 years B.P.)

As mentioned previously, the last appearance of significant amounts (greater than 35%) of authigenic dolomite in the sediments is correlated with the end of the mid-Holocene playa stage, which is placed at 5500 years B.P. From this point, the isotope record shows an overall decreasing $\delta^{18}O$ value with a minimum of -6.78 o/oo, equivalent to a maximum isotope lake-level of 1283 m (4209 ft) at 4600 years B.P., one meter below the historic high of 1283.8 m (4212 ft) in 1873 A. D. Afterwards, the isotope level shows a steady decrease until about 4000 years B.P. reaching a level below the modern average of 1280.8 m (4202 ft) but once again rises to higher stands equivalent or slightly above the modern average. The authigenic carbonate in the sediments is predominantly aragonite, but slight increases in percentage calcite corresponding to oxygen-18 depletion occur. We equate this early neolacustrine maxima-5 with the first maxima in the neopluvial episode (Currey and James, 1982). The isotope record indicates that mean lake levels during the early neolacustrine maxima-5 were as great as the modern mean and there were periods of wide fluctuations above and below this mean, similiar to the historic record. As suggested by Currey (1976), this early maxima represents a time of strengthened westerly flow and heavy winter precipitation supplying large amounts of isotopically light water to the lake, enough to decrease the salinity to the point where the percentage of calcite in the authigenic carbonate precipitate showed a notable increase.

Early Neolacustrine Minima-4 (3600-2900 years B.P.)

The isotope record for this relatively short period of 700 years shows a drop in lake level to as much as 5 m (15 ft) below the modern mean elevation. The average $\delta^{18}O$ value for this period is - 4.75 o/oo representing a mean isotope elevation of 1277 m (4190 ft), 4 m (12 ft) below the modern mean elevation. During this period increasing salinity must have affected carbonate precipitation as the percentage of aragonite rose to values greater than 35%. The isotope record indicates that the lake level was consistently lower during this period than at any other time in the late Holocene. It cannot be excluded that the summer precipitation is more enriched in oxygen-18 than the winter snowfalls, i.e. the fresh-water entering the lake has a more positive $\delta^{18}O$ value during a neolacustrine minima. This is most likely the case as the dominant source of the precipitation will vary between the maxima and minima periods, westerly flow versus Mexican Monsoons respectively. Whatever the isotope composition of the fresh-water input, the isotope record does show rather constant $\delta^{18}O$ values for the minima intervals without the wide fluctuations recorded for the maxima intervals. The minima intervals are furthered characterized by consistently more positive oxygen-isotope ratios. We interpret these observations as a reflection of a dominant evaporative control on the lake level during minima periods. Thus, it is the highly variable fresh-water input which regulates the widely fluctuating isotope values during the maxima periods.

Mid-Neolacustrine Maxima-3 (2900-1500 years B.P.)

The highest isotope elevation (1285 m) for the Great Salt Lake during the late Holocene was observed in the record associated with the mid-neolacustrine maxima-3. As mentioned previously, the salinity controlled changeover from aragonite to calcite precipitation was correlated with the high stand which resulted in the inundation of surrounding salt flats and marshes between 3500 and 2200 years B.P. (Mehringer, 1977). The minimum

$\delta^{18}O$ value of the calcitic sediments at this highest stand is -7.39 o/oo, which would be equivalent to a 3.9 m (13 ft) rise above the modern mean elevation. Our record for percentge aragonite shows a steady decrease from 38 to 14% beginning at about 2900 years B.P. and culminating with the change to calcite precipitation. Late Holocene beach ridges at about 1287 m (4222 ft) have been associated with this mid-neolacustrine high stand (Currey, 1977, 1980). The isotope lake-level of 1285 m (4217 ft) is, of course, a mixed value, as it represents isotope compositions for about a 20-year period or a sampling interval of 0.5 cm (approximately 0.26 mm/yr). The isotope record for the mid-neolacustrine high stand indicates that the lake elevation rose rapidly in less than 100 years to the highest isotope elevations observed for the late Holocene, remained at these high elevations for approximately 200 to 250 years, and descended steadily to levels below the modern mean elevation in approximately 200 years. The lake reached its peak height at about mid-way (700 years) into the mid-neolacustrine maxima-3. Apparently, as deduced from the isotope record, once climatic conditions for a transgressive phase are established the lake will rise quickly and maintain its high stand for a considerable period of hundreds of years and afterwards decline at a more gradual rate than the ascent. From our isotopic interpretations, we conclude that fresh-water input dominanted substantially over evaporative concentration. The exact nature of climatic conditions resulting in the dramatic increase in fresh-water accumulation in the lake, as indicated by the isotope and calcite composition of the sediments, are unfortunately not included in the sedimentary record. Perhaps, the rise is the result of a combination of factors reinforcing each other, strengthened Westerly flow with an increased snowfall plus a decrease in the summer evaporation of precipitation, run-off and surface waters of the lake.

Late Neolacustrine Minima-2 (1500-700 years B.P.)

During the latest 800-year neolacustrine minima-2, the isotope record indicates that the lake elevation fluctuations were smaller when compared with those of the neolacustrine maxima. The average δ^{18} O value is -5.75 o/oo representing an average isotope lake level at about 1279.9 m (4199 ft) or about 1 m below the modern mean elevation. The authigenic carbonate remained aragonite throughout this period indicating that salinity conditions conducive for calcite precipitation never developed. Neolacustrine minima are associated with increased summer rain fall and changes in vegetation patterns (Currey and James, 1982), but the isotope record between 1500 and 700 years B.P. indicates a consistent evaporative control on the lake level throughout the period. Apparently, summer precipitation reaches the lake as diminished pulses during a period when evaporative fractionation already dominates. This pattern of summer fresh-water input results in low-amplitude fluctuations of the isotope record correlatable to minima periods. The late neolacustrine minima-2 saw the development and expansion of indigenous peoples in the northeastern Great Basin probably as a direct consequence of the increased summer precipitation which would have facilitated both the hunting of grazing animals feeding on the increased grassland and the growing of abundant agriculture products. Therefore, it is undoubtedly not unrelated that the sudden disappearance of these peoples from the region corresponds with the termination of mesic climatic conditions and the onset of the late neolacustrine maxima-1.

Late Neolacustrine Maxima-1 (700 years B.P. to present)

Beginning about 700 years B.P. (1300 A.D.), the isotope record for the Great Salt Lake once again indicates a change in climatic conditions resulting in extreme and rapid fluctuations in isotope elevations characteristic of neolacustrine maxima. The isotope record for the past 700 years undoubtedly has a higher resolution due to less compaction of the sediments. As previously mentioned, the historic high stand of 1873 A.D. was correlated with the most negative of the more recent peaks in the isotope record (Figure 2). At least three other high stands, equal to or greater than the historic high, are notable in the pre-historic record within the late neolacustrine maxima-1 at approximately 1850 A.D., 1560 A.D. and 1400 A.D. The highest of these, 1560 A.D., has a $\delta^{18}O$ value of -7.03 o/oo which can be associated with an isotope elevation of 1284 m (4212 ft) or 3 m (9 ft) above the modern mean elevation. The carbonate mineralogy data show a slight increase in percentage calcite with a sharp decrease in percentage aragonite in the sediments corresponding to this period of decreasing salinity or rise in lake level. Geomorphology studies indicate that there was a late pre-historic high stand culminating at 1285 m (4217 ft) around 1700 A.D. (Currey, 1980; Currey et al., 1984). Considering that the isotope signals represent about 20-year intervals and, hence, are mixed values, we propose that the 1560 A.D. maximum-1 isotope elevation during the late neolacustrine maxima has a topographic manifestation corresponding to the 1285 m high stand. The approximate stratigraphic date for the highest stand of the late neolacustrine maxima-1 corresponds to the beginning of a climatic deterioration in Europe known as the "Little Ice Age". In the northeastern Great Basin region of Utah, this deterioration is marked by rapid changes in the elevation of the Great Salt Lake, probably a result of periodic increases in the fresh-water input from increased winter precipitation.

CONCLUSIONS AND PREDICTIONS

Using a combination of high resolution oxygen-isotope stratigraphy and percentage carbonate and carbonate mineralogy in samples from sediment cores, a detailed, late Holocene record of lake-level fluctuations has been compiled for the Great Salt Lake, Utah. The division of the isotope record into 3 maxima and 2 minima intervals within the neolacustrine episode beginning about 5500 years B.P and continuing until the present is in total agreement with the other paleoclimatic interpretations for the northeastern Great Basin (Currey and James, 1982). According to the climatic scenario of Currey (1976), we associate the isotope minima (oxygen-18 depletion) with periods when westerly flow and winter precipitation dominated and the isotope maxima (oxygen-18 enrichment) with strengthened Mexican Monsoon circulation and summer precipitation. Our approximated stratigraphic ages for the sedimentary record indicate that the neolacustrine maxima have a duration of about 1500 years while minima span about 750 years.

The isotope record indicates that the northeastern Great Basin is approximately 700 years into the most recent maxima and that extreme fluctuations between high and low stands around a mean modern elevation are to be expected. A conservative interpretation of the isotope record predicts that the Great Salt Lake could once again steeply rise to the highest topographic elevation of 1285 m (4217 ft) recorded during the present neolacustrine maxima-1. This would be a short-lived high stand with modern mean elevations again obtained within 50 years. A broader interpretation of the isotope record, in particular using the data of the mid-neolacustrine maxima-3, predicts that the Great Salt Lake could reach higher topographic elevations of up to 1287 m (4222 ft). The isotope record for the mid-neolacusterine maxima-3 indicates that the highest elevation occurred at about the mid-point of the maxima period and the duration of the rise, high stand and return to a mean modern elevation spanned approximately 500 years, which can be divided into 100, 200-250 and 200 year segments, respectively. The steep rise in the lake elevation since 1983 A.D. to levels near or equal to the historic high stand at 1284 m (4212 ft) strongly suggest that climatic conditions prevailing during a neolacustrine maxima continue to dominate and even higher elevations, as recorded in the isotope stratigraphy, can be anticipated.

ACKNOWLEDGEMENTS

We would like to acknowledge the numerous colleagues with whom we have worked on the Great Salt Lake. In particular, we thank Q. He and J. Pika, for their assistance with the laboratory analysis, S. Joyce for technical assistance, and D. Currey for his helpful discussions on stratigraphic correlations in the Great Salt Lake Basin. This is contribution No. 274 of the Laboratory for Experimental Geology, ETH-Zurich.

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Table 1: Historic Lake Elevations versus Oxygen-Isotope Values

Year (A.D.)	$\delta^{18}O_{PDB}$ (°/oo)		$\Delta\delta^{18}O_{PDB}$ (°/oo)		h(m)	Δ h(m)	$\lvert\Delta\delta^{18}O/\Delta h\rvert$ (°/oo/m)	
	gravity	piston	gravity	piston			gravity	piston
1978	-6.17		-0.93		1280.2	+2.8	0.33	
1963	-5.24		+1.06		1277.4	-3.1	0.34	
1952	-6.30	-6.28	-0.43	-1.20	1280.5	+2.2	0.20	0.54
1933	-5.87	-5.08	+1.11	+1.61	1278.3	-3.4	0.33	0.47
1925	-6.98	-6.69			1281.7			
1910		-6.30			1281.4	+2.5		0.48
1905	-5.70	-5.11		-1.19	1278.9	-4.9	0.30	0.30
1873	-7.18	-6.57	+1.48	+1.46	1283.8	+3.9	0.29	0.22
1862	-6.04	-5.72	-1.14	-0.85	1279.9			
Average Modern 1851-1978 A.D.	-6.06 o/oo				1280.8 m			0.35 o/oo/m

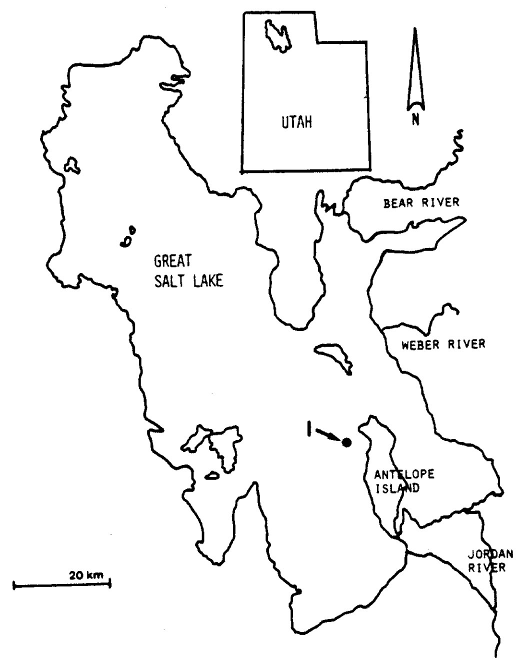

Figure 1. Map of Great Salt Lake, Utah showing location of Core Site I.

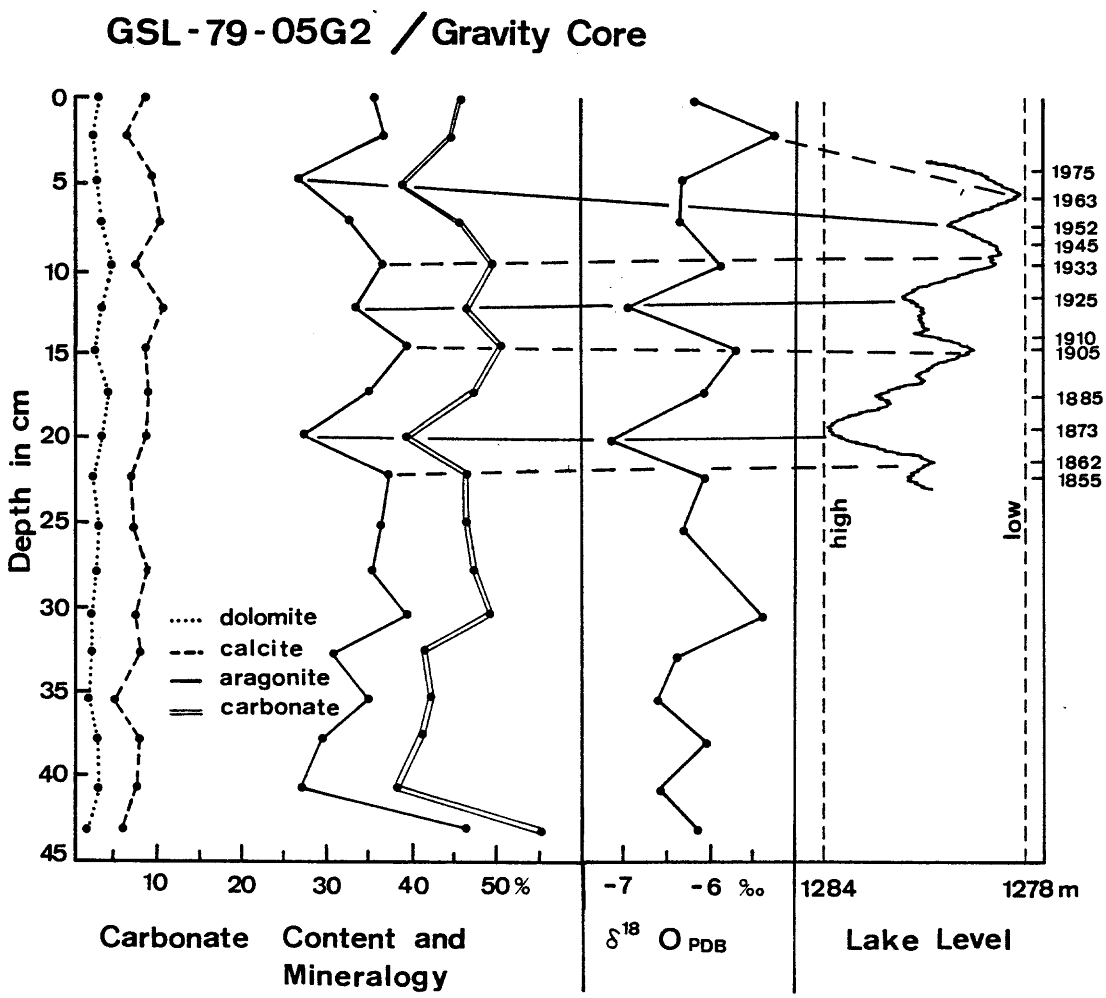

Figure 2. Carbonate and isotope stratigraphies correlated with historic lake-level curve (after Austin, 1980). See text for discussion.

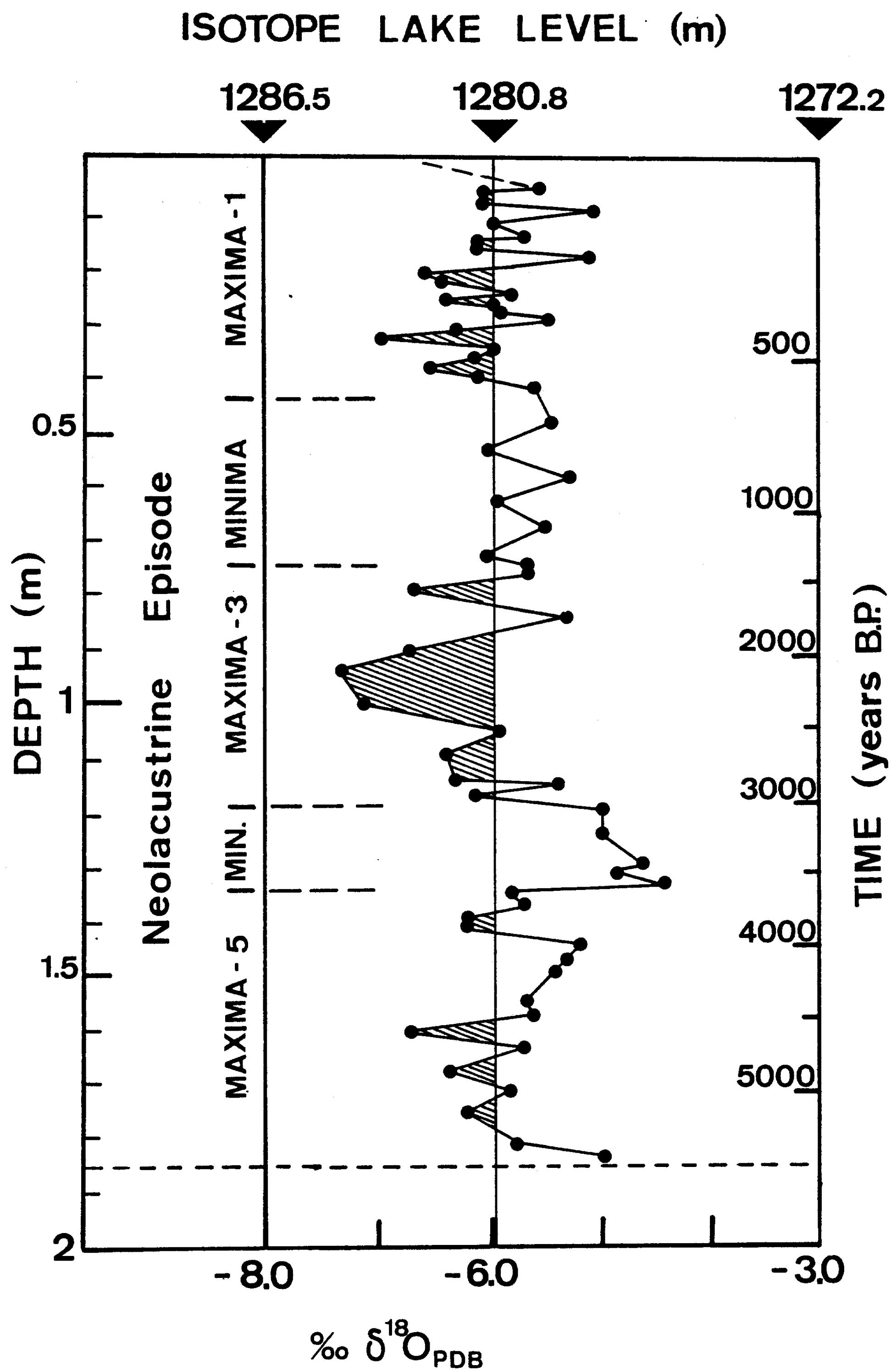

Figure 3. Oxygen-isotope stratigraphy for upper 2 m of piston core (GSL-79-05P2) showing correlation of data with calculated isotope lake-levels on an approximated time scale. Maxima periods with decreased δ^{18} values represent times when the lake level shows extreme fluctuations frequently rising above the mean modern elevation (1280.8 m). Minima periods with increased δ^{18} values correspond to times when the lake level falls and remains below the modern mean elevation. See text for discussion.

APPLICATION OF AN ENERGY-BALANCE MODEL TO THE LATE PLEISTOCENE LITTLE COTTONWOOD CANYON GLACIER WITH IMPLICATIONS REGARDING THE PALEOHYDROLOGY OF LAKE BONNEVILLE

William D. McCoy, Department of Geology and Geography, University of Massachusetts, Amherst, MA 01003

Larry D. Williams, Department of Geography, University of Aberdeen, Aberdeen, Scotland, U.K.

INTRODUCTION

The cause of increased lake levels in closed basins of the western United States during the late Pleistocene has been a matter of controversy for many years. Some workers contend that the time of lake-level maxima was relatively cold and dry in terms of mean annual temperature and precipitation. High lake levels were the result of increased runoff due to reduced evapotranspiration and decreased lake evaporation (Galloway, 1970; Brakenridge, 1978). Regional precipitation rates may have been similar to, or even somewhat less than, present rates. Some other workers believe that the late Pleistocene temperature reductions that they infer could not sufficiently reduce evapotranspiration and lake evaporation to maintain the increased lake levels unless a concomitant increase in precipitation occurred (Leopold 1951; Antevs, 1952; Snyder and Langbein, 1962).

The notion of balancing the water budgets of lakes in terms of a change in mean annual temperature versus a change in the annual precipitation is an obvious oversimplification of a complex environmental system. Certainly many factors besides temperature are responsible for evapotranspiration and lake evaporation. In addition, the seasonality as well as the total amount of precipitation is important in determining the amount of runoff produced. Nevertheless, the maximum size of many Pleistocene closed-basin lakes may be used to constrain estimates of the annual precipitation change that took place in the basin in combination with any given temperature change, if conservative values are chosen for other important parameters such as mean wind velocity, average relative humidity, and extent and type of cloud cover. Obviously, an infinite set of combinations of change in mean annual temperature and annual precipitation values are possible.

However, in the case of Lake Bonneville at the time of its maximum level (fig. 1), there is no constraint on the maximum annual precipitation for any given change in temperature because the lake overflowed and lost an unknown volume of water annually to the Snake River drainage. An infinite set of combinations of temperature and precipitation changes exist that would maintain the lake at its maximum level, but the uncertainties of the water budget of the lake

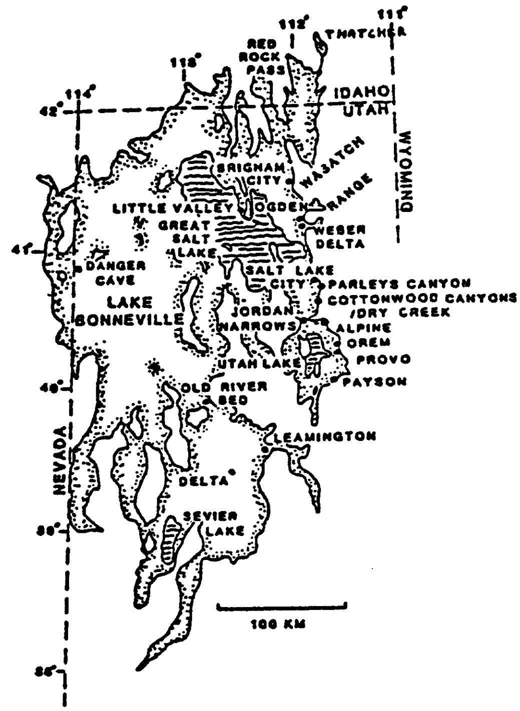

Figure 1. Map of the extent of Lake Bonneville at its maximum. Little Cottonwood Canyon is located on the eastern edge of the lake and is labeled Cottonwood Canyons (modified from Scott, 1980).

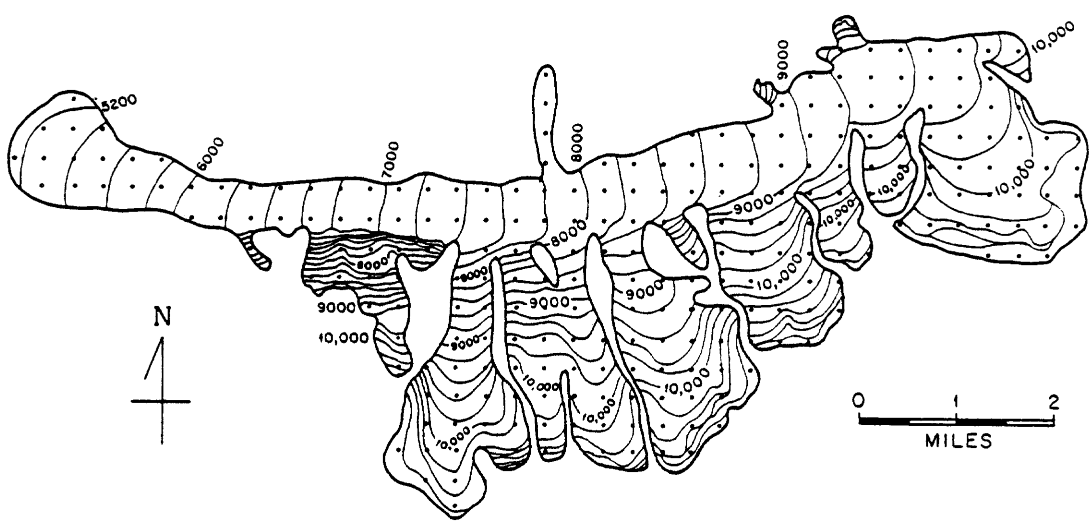

Figure 2. Reconstruction of glacier surface in Little Cottonwood Canyon (from Gallie, 1978) with location of grid points used for numerical reconstruction of surface in this study.

can provide no upper limit to the amount of precipitation change.

A large valley glacier occupied Little Cottonwood Canyon (fig. 1 and 2) as Lake Bonneville rose to its maximum level. The mass balance of that glacier must have responded to the same climatic changes that caused Lake Bonneville to rise. Although the position of the terminus of Little Cottonwood glacier at the time of the Lake Bonneville maximum (c. 14,500 yr B.P.) is not known, it can be shown stratigraphically that the glacier maximum preceded the maximum lake level and that the ice was at less than its maximum extent at that time (Morrison, 1965a; 1965b; Richmond, 1964; Scott and others, 1983). If the mass balance of the Little Cottonwood glacier can be modelled, it would be possible to estimate the maximum local precipitation corresponding to a given temperature change that will be just sufficient to maintain the glacier near its maximum position. This paper reports the results of an attempt to make such an estimate using an energy-balance snowmelt model to simulate the mass balance of the late Pleistocene glacier in Little Cottonwood Canyon.

THE MODEL

The surface of the Little Cottonwood Canyon glacier at its maximum extent is represented in the net mass balance model by a total of 253 grid points (fig. 2). The horizontal and vertical coordinates of each point were determined using a reconstruction of the glacier surface by Gallie (1978). The spacing between each grid point is 483 m in both the north-south and east-west directions. The simulation of the net mass balance of the Little Cottonwood glacier utilizes a calculation of the energy budget of the glacier surface. The hourly energy balance calculations follow a modification of a net mass balance model described by Williams (1974; 1975). The energy balance takes into account the net shortwave and net longwave radiation and the sensible and latent heat fluxes at each grid point. Conduction from the ground is neglected. Air temperature at each grid point is determined using a lapse rate calculated from mean monthly data from Cottonwood Weir and Silver Lake Brighton (2652 masl) for the period 1951-1960 (from U.S. Dept. of Commerce, Climatography of the U.S., No. 86-4). The daily temperature progression is assumed to be a sinusoidal function of time with a maximum at 1500h and a minimum at 0300h and with an amplitude determined by interpolation between the above-named stations. Mean monthly values of wind speed, cloud cover, and dewpoint temperature were taken from the records at the Salt Lake City airport, the nearest station with the available records. These values are not satisfactory as absolute values, but their relative values are considered to be useful. The sensitivity of the model to changes in wind speed, cloudiness and dewpoint temperature is discussed below.

Precipitation in the canyon is determined as a function of elevation at each grid point using the gradient established by the mean monthly values at Cottonwood Weir and Silver Lake Brighton for the period 1951-1960. The proportion of precipitation falling as snow is determined as a function of the mean monthly air temperature at each grid point. The function was found by regressing the mean monthly percentage of precipitation that falls as snow against mean monthly temperature at Silver Lake Brighton and Cottonwood Weir (fig. 3). The relationship is similar to that reported for other high altitude stations (Barry, 1981, p. 191).

Avalanching of snow onto the glacier was probably a significant source of mass accumulation. Avalanching is neglected in this model and, as a result, the simulations probably predict slightly more precipitation than would be necessary to maintain the glacier's equilibrium.

TEST OF THE MODEL

The model's output of monthly snow accumulation and melt can be compared with snow course data as a test of the computer model. Data for the water equivalent of the snowpack at Silver Lake Brighton is available for the first of each month from January to June (Soil Conservation Service, 1979). Using averages of climatic data for the period 1951-1960 as input for the model, a comparison of predicted water equivalent with the average measured values for the same period reveals less than a 10% error in the predicted values for the first of each month (fig. 4). For June 1 a direct comparison cannot be made because in many years the snowpack had completely melted before June 1 and the total amount of snow that would have melted by June 1 is not known. Those potentially negative values of water equivalent are not available to calculate a comparable average for June 1. Therefore, the "measured" value plotted in figure 4 is an overestimation of the true average.

The slight but consistent underestimation of snow accumulation at Silver Lake is interesting. In the cold months of December, January, and February when there is virtually no melting of the snowpack the generation of snow by the model is slightly less than the average of the measured values. However, the input for the model uses the 1951-1960 Silver Lake precipitation averages precipitation gage measurements. The discrepancy between the precipitation values used for input in the model and the amount of snow measured on the ground appears to be an artefact of the methods of measurement. The discrepancy (3-9%) appears to be the same magnitude and in the same direction as the error attributed to the use of the standard Federal snow sampler (Work and others, 1965; Farnes and others, 1983).

Computed snowpack water equivalent was also compared with average snow course data from Mill D South Fork for the same period. The model overestimates the amount of snowmelt at that site. The discrepancy between the actual and

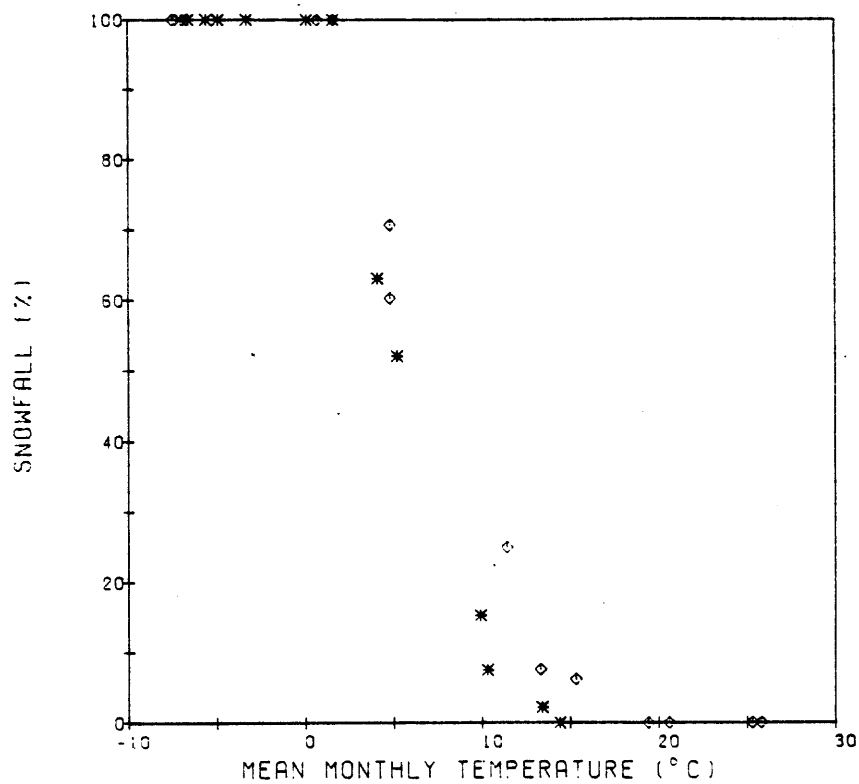

Figure 3. Plot of percentage of mean monthly precipitation that falls as snow versus mean monthly temperature (for period 1951 - 1960). Asterisks refer to data from Silver Lake Brighton and diamonds represent values from Cottonwood Weir.

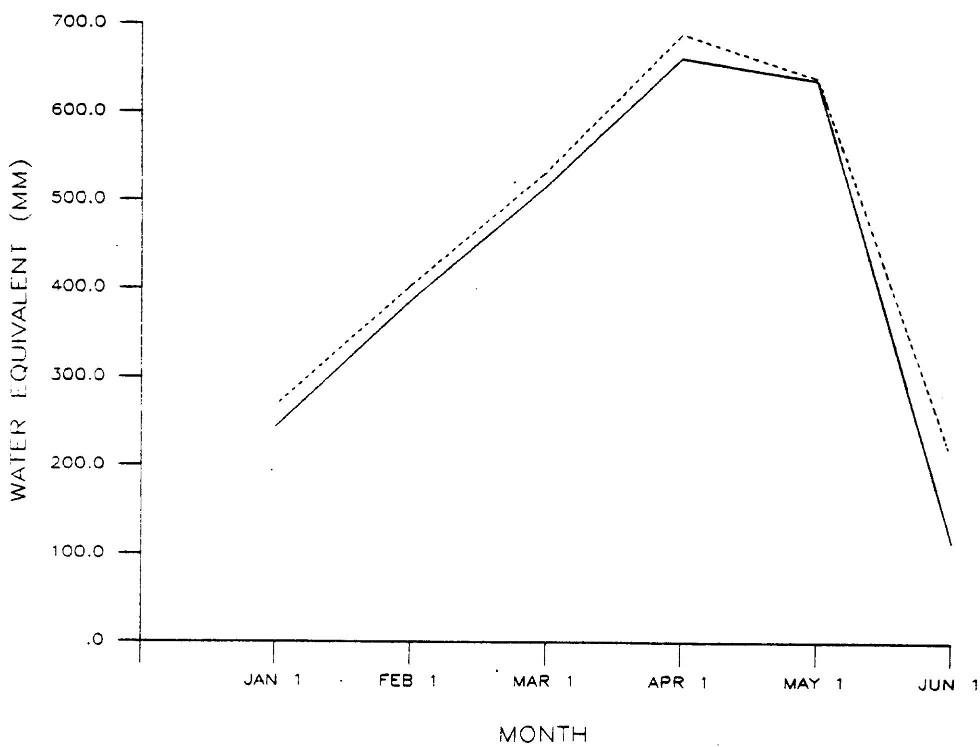

Figure 4. Comparison of measured (dashed line) versus simulated (solid line) mean water content of snowpack at Silver Lake Brighton (mean for period 1951 - 1960).

predicted snowmelt at Mill D South Fork is roughly the amount that would be expected given the amount of vegetation (aspen) cover at that site. The reduction of net radiation caused by increasing canopy density has been summarized by Dunne and Leopold (1978, p. 474).

APPLICATION OF THE MODEL TO THE LITTLE COTTONWOOD CANYON GLACIER

The results of running the net mass balance model are expressed in terms of the change in mean annual temperature versus the change in precipitation (fig. 5). In figure 5, it is assumed that each mean monthly temperature was reduced by an equal amount and that the monthly proportions of precipitation are the same as at present. The change in precipitation is given as the multiple of the present precipitation rate that yields sufficient snow to just balance accumulation and ablation on the Little Cottonwood Canyon glacier at its maximum extent. A mean annual temperature depression of 4.8°C would require about 2.5 times the present annual precipitation for the glacier to be maintained at equilibrium while a 16°C decrease in temperature would need only about 0.75 times the present precipitation to sustain the ice at its maximum extent. Representative net-balance curves from model simulations for three different combinations of temperature and precipitation changes that maintain the glacier at equilibrium at its maximum extent are shown in figure 6. The equilibrium-line altitudes are the same for all three simulations, but the wide variation in mass flux is evident. A small temperature reduction (3.6°C) combined with a large increase in precipitation (2.853 times the present) produces a glacier with a steep net-balance gradient, a high mass flux, and a high activity index. A large temperature reduction with virtually no change in annual precipitation produces a glacier with a lesser net-balance gradient, a lower mass flux, and a lower activity index.

Galloway (1970, 1983) and Brakenridge (1978) present evidence supporting "full-glacial" temperature depression of 7-10°C in the southwestern United States. The time of maximum extent of glaciers in that region is very poorly known, however their temperature estimates may prove to be approximately correct for the time of the potential maximum stand of Lake Bonneville. McCoy (1981) calculated a minimum temperature depression of 7°C for the northeastern Great Basin for the interval 16,000-11,000 yr BP based on the rate of epimerization of isoleucine in fossil gastropod shells. Thompson and Mead (1982) also report evidence for a cold, dry late Pleistocene climate in the vicinity of the Snake Range in the north central Great Basin. Using these estimates of temperature depression for the time of the potential maximum of Lake Bonneville, a precipitation rate of about 175% of present would maintain the Little Cottonwood glacier at its

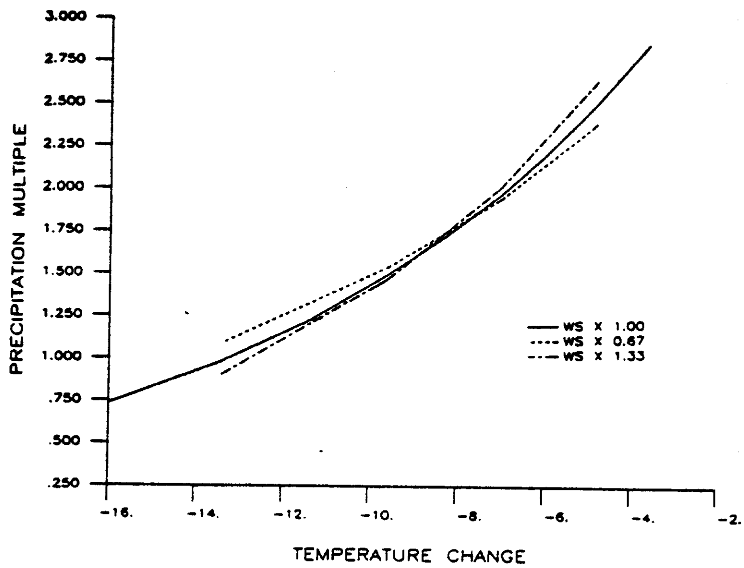

Figure 5. Plot of temperature change versus precipitation change that maintains a glacier in equilibrium at its maximum extent in Little Cottonwood Canyon. Dashed line and dash-dot line demonstrate the effects of changes in mean monthly wind speeds on the net mass balance simulation in terms of the required multiple of present precipitation needed to maintain the glacier at a given temperature reduction.

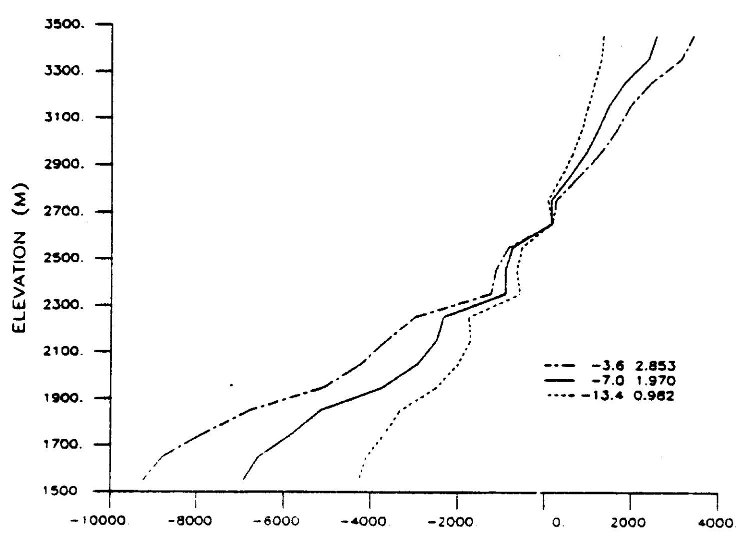

Figure 6. Plot of net balance versus elevation for three different possible climatic regimes in Little Cottonwood Canyon. The dash-dot line represents a situation with monthly temperatures 3.6°C below present and monthly precipitation values 2.853 times those of the present. The solid line is for temperatures 7.0°C below present and precipitation 1.970 times present values. The dashed line is for temperatures 13.4°C below present and precipitation 0.982 times that of today.

maximum size. Since the glacier was at some unknown, but lesser, extent at that time, this estimate of the precipitation rate is a maximum given the assumptions concerning other climatic variables discussed above. The following section discusses the effects that changes in assumed mean monthly wind speed, dewpoint temperature, and cloudiness might have on the net mass balance of the Little Cottonwood glacier. And, perhaps more important, the effects of changes in the seasonal distribution of temperature and precipitation changes are discussed.

The effects of changes in windspeed can be seen in figure 5. A 33% increase in mean monthly windspeeds results in the need for increased precipitation to maintain the glacier at mean annual temperature depressions of the less than 7 or 8°C. The model shows an increased sensible heat flux to the ice that results in increased ablation. With mean annual temperature depressions of more than about 8°C the windspeed increases have little effect because the mean monthly summer air temperatures are too low to advect much sensible heat and the loss of latent heat with evaporation becomes more important. This latent heat loss becomes dominant over sensible heat gain with large temperature depressions. A windspeed decrease of 33% causes somewhat less summer sensible heat transfer to the ice and results in a slightly reduced precipitation multiple necessary to maintain the glacier at temperature depressions in the range of 5 - 9°C. The importance of the model experiments with windspeed changes is that in the probable range of temperature depression appropriate for the time of Lake Bonneville maximum (7 - 10°C) changes in windspeed have very little effect on the amount of ablation because of low summer air temperatures. In this temperature range, it is possible that increases in windspeed would do more to enhance the net mass balance by increasing accumulation on the glacier by snow redistribution than it would to reduce the net mass balance by advecting heat.

Changes in the mean monthly dewpoint temperatures have relatively little effect on the simulated mass balance until the point is reached where the mean dewpoint temperature is close to the mean monthly temperature. At that point the air is saturated with water vapor much of the time and a large flux of latent heat to the glacier surface occurs during the summer months. In the simulations presented here, the mean monthly dewpoint temperatures are equal to the present mean monthly dewpoint temperature minus one-half of the reduction applied to the monthly temperatures.

The effect of changes in mean monthly cloud cover are shown in figure 7. Although it is well known that the type of cloud is as important as the extent of cloud cover in determining the net radiation at the surface, the model does not account for cloud type and no data on the frequency and extent of cover of clouds of different types in this area were available to the author. However, the effect of changes in the general extent of cloud cover were examined (fig. 7).

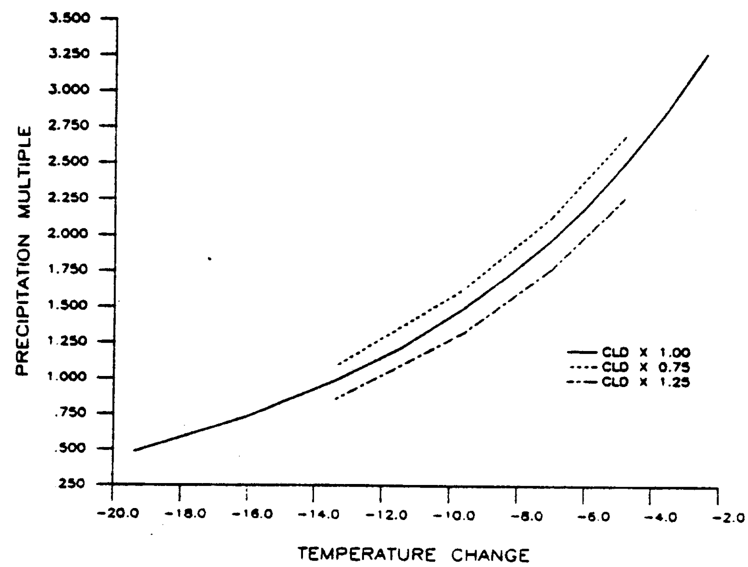

Figure 7. The effects of cloudiness on the Little Cottonwood glacier expressed in terms of the precipitation required to maintain the ice at its maximum extent with a given reduction in monthly temperatures.

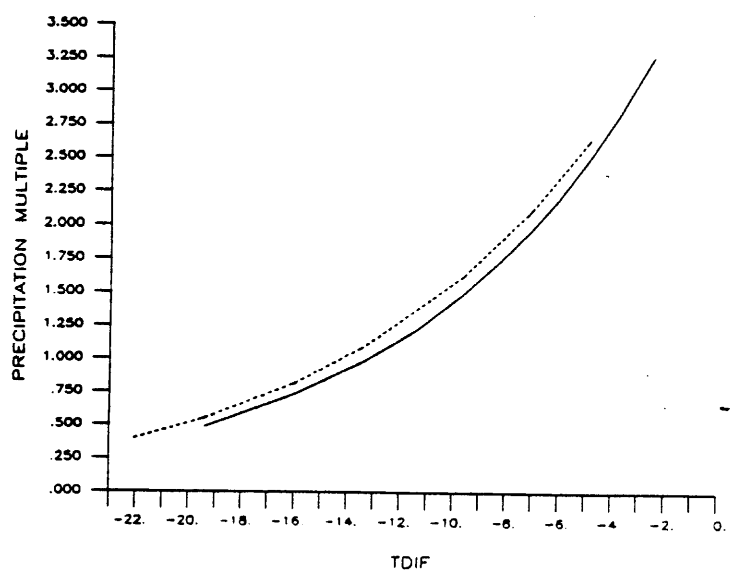

Figure 8. The effects of an altered temperature regime (dashed line) on the Little Cottonwood glacier. Values on the abscissa refer to changes in the temperature of the warmest month (July). Reductions in temperatures for other months are less, being proportional to the sine of (K-1)PI/12, where K is the number of the month and PI is the number of radians in 1/2 of a circle. The solid line represents values of the unaltered temperature regime.

An increase in average cloud cover of 25% for each month resulted in a net decrease in the precipitation necessary to maintain the glacier at a given temperature depression. The increase in cloud cover increases the net longwave radiation at the glacier surface but is more effective in reducing the amount of shortwave radiation received at the surface. Therefore, the net radiation and the energy available for melt are reduced with an increase in cloud cover. The 25% increase in cloud cover results in a 9 - 13% reduction in the precipitation needed to maintain the glacier. Similarly, a 25% reduction in cloudiness for each month results in a 7 - 10% increase in the amount of precipitation necessary to sustain the glacier.

Several workers have argued, though not convincingly, that Pleistocene temperature reductions would have been greatest in the summer and might have been negligible during the winter months. The effects of such a temperature regime on the Little Cottonwood glacier model are shown in figure 8. The reduction of temperatures in this model experiment follow a sine function such that the warmest month, July, experiences the full temperature reduction as indicated on the abscissa of figure 8, and the mean maximum and minimum temperatures of the coldest month, January, are not reduced at all. Intermediate months experience temperature reductions proportional to the sine of (K-1)PI/12, where K is the number of the month. Therefore, the reduction in mean annual temperature is equal to 0.633 times the maximum monthly temperature reduction as indicated on the abscissa of figure 8.

The experiment indicates that the precipitation required to maintain the Little Cottonwood glacier in the situation where the summer temperatures are reduced most is not much more than the precipitation required when the same temperature reduction is applied to all months. This result is not surprising since nearly all ablation takes place in the summer months. Reducing winter temperatures has little effect on the net mass balance of the glacier.

If we apply this altered temperature regime and hold all other factors constant, it can be seen (fig. 8) that if the July temperature reduction is, for example, about 13°C, the reduction of the mean annual temperature is only about 8°C (0.633 times 13°C) and a precipitation rate equal to that of the present is sufficient to maintain the glacier. A precipitation rate of about 175% of the present rate is needed to maintain the glacier if the same mean annual temperature reduction is produced by an 8°C reduction of all monthly temperatures throughout the year. Therefore, knowledge of the summer temperature reduction at the time of lake maximum is important for the modelling of the paleohydrology of the Bonneville basin.

It is not unlikely that the seasonality of precipitation was altered at the time of the maximum stand of Lake Bonneville. At present the precipitation in the Wasatch Mountains reaches maximum values during the coldest half of the year. The net mass balance model was used to examine the effects of altering the precipitation regime (fig. 9). In this experiment the precipitation for the coldest six months (November - April) was reduced by 50%. An equal amount of precipitation was evenly distributed among the six warmest months, resulting in a strong, warm-season precipitation maximum. The results indicate that this drastic change in precipitation regime actually reduces the precipitation multiple that is required to maintain the glacier at a given temperature reduction. The only exceptions are when the temperature reductions are very small (less than 3°C). The reason for this effect is that with temperature depressions of more than about 3°C most of the precipitation at higher altitudes falls as snow, even during most of the six warmest months. Although the total annual snowfall is not increased, large amounts of fresh snow during the ablation season tend to maintain a higher average albedo and thereby reduce the amount of shortwave radiation absorbed at the surface. The rain that does fall does not contain sufficient heat to melt a significant amount of ice over a large portion of the glacier.

CONCLUSIONS

A physically based model of the net mass balance of a glacier can provide insights to the paleoclimatic implications of glacier fluctuations. In the case of the Little Cottonwood glacier, this model suggests that if the mean annual temperature (or even only the mean summer temperature) was more than 7 or 8°C less than present, the annual precipitation in Little Cottonwood Canyon at the time of the maximum stand of Lake Bonneville could have been no more than about 175% of the present value. Given the fact that the Little Cottonwood glacier was not at its maximum extent at that time and may have receded significantly by then, this maximum precipitation value may be much greater than the true value. Furthermore, because the entire glacier lay within a few miles of a large, deep lake, a significant local lake-effect enhancement of precipitation must have occurred, especially in the autumn and early winter months. Braham and Dungey (1984) have shown a present-day lake-effect enhancement of snowfall in the lee of Lake Michigan ranging from 30 to 120%. The Wasatch Mountains to the lee of Lake Bonneville may have served to make the lake-effect enhancement of snowfall even more effective during the high stands of that lake. Therefore, even though the late Pleistocene local annual precipitation in Little Cottonwood Canyon may have been as much as 175% of the present value, the regional annual precipitation may not have been much, if any, greater than the present value. That is to say that the annual precipitation over the whole of the Bonneville basin at the time of the highest stand of the lake may not have been much different than the present. Locally, along the

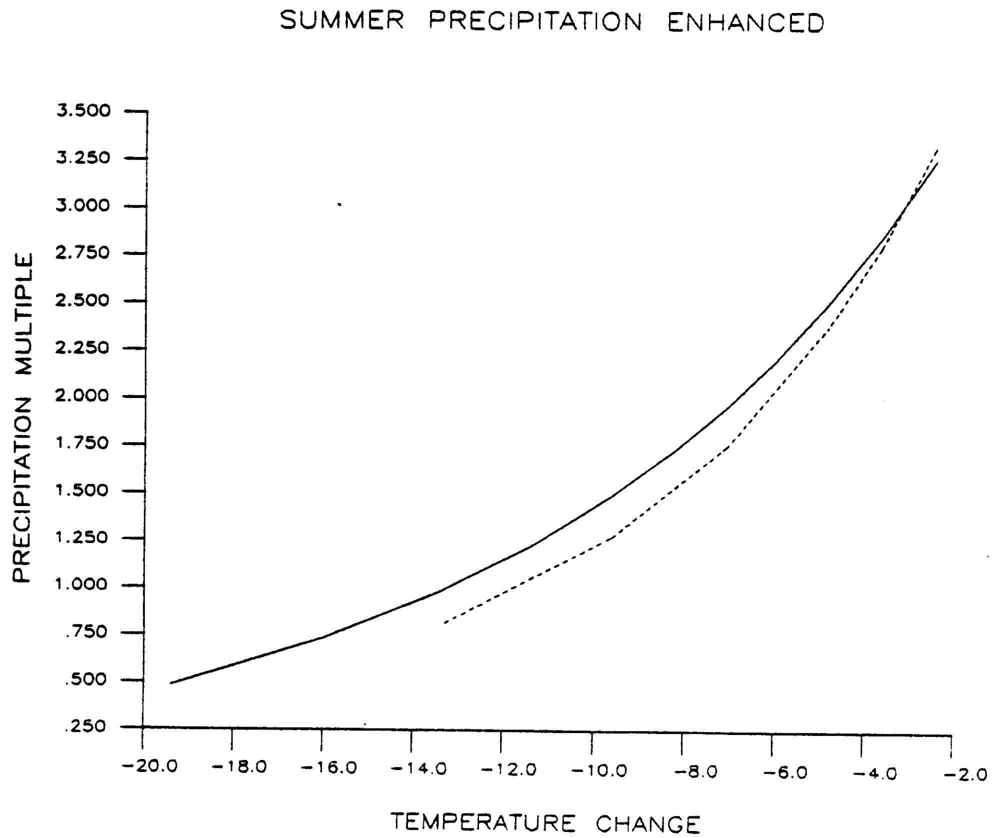

Figure 9. The effects of an altered precipitation regime (dashed line) on the Little Cottonwood glacier. Precipitation values for the six winter months (November - April) are reduced by 50% and an equal amount is apportioned evenly among the six summer months. The resulting precipitation regime has a strong summer maximum of precipitation which enhances the net balance of the glacier mostly because of the resultant increase in mean summer albedo. The solid line represents values of the unaltered precipitation regime.

Wasatch Front in particular, the annual precipitation may have been much greater than today as fall and early winter precipitation must have been greatly enhanced by the lake.

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CORRELATION BETWEEN VEGETATIVE AND LAKE LEVEL CHANGES: The Pollen Evidence

David B. Madsen

It has become increasingly clear that paleoclimates can be interpreted from a cline of decreasingly useful proxy data ranging, in a general way, from geomorphological information, to data on the distribution and composition of plant communities, to information from the animals dependent on those plants. This is less of a problem when interpreting information derived from marine environments since, in terms of micropaleontological data, the cline is a comparatively narrow one. When interpreting terrestrial proxy data, on the other hand, the cline is relatively broad and the comparative usefulness of these various kinds of information is marked.

For a similar reason, the comparability of these data is often limited because they react so differentially, in terms of both the nature and rate of change, to climatic forcing mechanisms. In the Great Basin, for example, it has become abundantly clear that geological features such as pluvial lakes and mountain glaciers are much more sensitive to climatic changes associated with the close of the last Pluvial than are the distribution and density of various plants and animals. During that period, mountain glaciers and closed-basin lakes had, within a period of roughly a thousand years, reached the level of their Holocene average, while plant communities underwent a series of complex changes in both composition and distribution and did not reach their modern analogs until 7-8000 years ago or even more recently. What this means, quite simply, is that in attempting to predict future levels of Great Salt Lake on the basis of medium to short-term fluctuations, it is probably more appropriate to use physical data such as geomorphological features or mineralogical/isotope changes to identify the pattern of past events than it is to use less sensitive proxy data such as the nature and distribution of past plant communities.

There is a further problem of scale when dealing with the interpretation of past plant communities. Plant macrofossils from woodrat nests, bog deposits and other preservational mediums can usually be identified to the species level and, because of the limited tolerances of many plant types, can provide excellent proxy data for local climatic conditions. However, these macrofossil records essentially represent point localities in both time and space, and, given the vagaries of local environmental conditions (especially in an area as diverse as the Great Basin), become less useful as more generalized paleoenvironmental interpretations are required. The situation is virtually opposite when dealing with plant microfossil data such as fossil pollen and spores. With a few important exceptions, most pollen types cannot be identified with any degree of reliability to more than genera and, hence, present comparatively broad environmental tolerances with which to interprete the fossil record. On the other hand, pollen in depositional sites like the Great Salt Lake is transported great distances and such pollen records provide a general paleovegetational record of a relatively large area. While a few closely dated pollen sequences from sensitive ecotone sites provide more detailed vegetational records, the most optimal situation is clearly one in which both pollen and plant macrofossil data are available for interpretation. Plant macrofossil data is comparatively more useful in situations, such as the very Late Pleistocene and Holocene, where the number of samples is large and they are spread over a relatively broad area. Pollen data is comparatively more useful in situations where macrofossil records are spotty or non-existent.

It would seem then, that the utility of pollen analysis for identifying short to medium-term climatic fluctuations is relatively limited when compared to other available interpretive tools. With increasing time depth, however, both plant macrofossil data and physical evidence become more rare. Macrofossil information is more limited because of increasing preservational problems and physical evidence is limited because it is destroyed or obscured by later features. Pollen data, on the other hand, becomes comparatively more useful with increasing time depth, if only because their utility remains constant while that of many other information sources decreases relatively rapidly.

In areas where depositional and preservational media are sufficiently good, such as the closed-basin lakes of the western U.S., pollen analysis is one of the more valuable interpretive tools for understanding long-term climatic fluctuations. In conjunction with minerological and microfaunal data from cores of lake deposits which span the last 3 million years or more (e.g., Eardley et al. 1973; Smith 1979)), pollen data can help determine the climatic regimes under which various lake depositions occurred. Pollen data are particularly important because, unlike minerological and microfaunal data, they are not controlled by lake levels and can be used as independent climatic proxy data. Since a variety of non-climatic geologic events, such as the capture of streams like the Bear River or catastrophic failure of bedrock dams like Red Rock Pass, can be responsible for changes in lake elevation, an independent guide to climatic conditions is critical to the interpretation of lake cores. Since it is becoming evident that the mid-latitude continental climatic record may not be strictly comparable to the Quaternary marine isotope record (Smith 1984; Bischoff et al. 1985), the correct interpretation of these lake cores is of increasing paleoenvironmental importance. If, in fact, the marine and continental records are not comparable, then pollen and other similar data from these deep closed-basin lakes may be one of the few available guides to continental climates on time-scales equivilent to Milankovitch cycles. However, for these pollen core records to be useful as guides on the long-term, it is necessary to determine how closely they parallel other evidence of climatic change on the short-term.

Non-lake Holocene Records

A limited number of short-term pollen records are available from the area of the Bonneville Basin. Relatively continuous records are available from Curelom Cirque in the Raft River Mountains north of the Great Salt Lake (Mehringer et al. 1971; Mehringer 1977), Snowbird Bog in the Wasatch Mountains east of the lake (Madsen and Currey 1979), and Swan Lake on the extreme northern end of the basin (Bright 1966). A larger number of records, which are less useful because they are discontinuous, poorly published, work in progress, or derived from culturally contaminated archeological sites, supplement these more continuous records.

The early portions of these Holocene records do not correlate well with the physical evidence of comparatively rapid deglaciation and lake level decline. Both Curelom Cirque and Snowbird Bog became ice free prior to about 12,500 years ago, while the area around Lowder Creek Bog, near Brianhead more than 4° further south, probably became ice-free some 1-3000 years earlier (L.W. Lindsay 1985, personal communication). The level of the Great Salt Lake had receded to below about 1313m (4310ft) by about 11,000 years ago, possibly reaching modern levels of 1280m (4200ft) or below by that time (Currey and James 1982). This evidence of relatively rapid glacial and lake level change is not reflected in pollen from surrounding bog and lake deposits. At Curelom Cirque, Snowbird Bog, and Swan Lake, the transition from Pluvial vegetation occurred over a period of 3-5000 years and did not reach anything comparable to modern conditions until 8-9000 years ago. Specific components of modern vegetation, such as pinyon pine, Utah juniper, and joint fir appear differentially in different areas of the eastern and central Great Basin as much as 2-4000 years later than that (Thompson 1984; Madsen 1985).

This delay in the transition from Pluvial to post-pluvial plant communities may be related as much to the magnitude of change as it is to a lack of climatic sensitivity. Much of this change included an altering of the composition of many plant communities as some species invaded their modern range from Pleistocene refugia. Since the mechanisms involved in the migration of these various species were apparently different, it is not surprising that they reached their modern limits at different times during the post-Pluvial period. As a result, during the Pluvial/post-Pluvial transition period, it is difficult to distinguish between short to moderate-term, smaller magnitude, climatically induced vegetation changes and change resulting from slow acting biological mechanisms associated with the greater magnitude climatic change of several thousand years earlier. However, once major plant communities became established by the early mid post-pluvial, they become more sensitive, in terms of pollen changes at least, to climatically induced changes simply because it is possible to detect changes in the relative proportion of plant types within those communities without the distraction of changing compositional elements.

Mid-to-late Holocene pollen samples from bog and marsh deposits around the periphery of the Great Salt Lake are sensitive enough to detect climatic changes on the order of the Altithermal or Neoglacial episodes. These medium-term climatic fluctuations are reflected in elevational changes of the Great Salt Lake on the order of about 15-20m. Evidence of warmer mid-Holocene conditions between about 7500 and 5000 B.P. when the Great Salt Lake virtually dried up (Currey and James 1982), is present in the both the Snowbird Bog and Curelom Cirque records, but is less clear in the Swan Lake sequence. Shorter duration pollen records to the west and southeast of the lake reflect a similar degree of change (Lindsay 1980; Thompson 1984). Evidence of a two-phase Altithermal period similar to that defined in the Rocky Mountains is less clear cut, although at Snowbird bog, temperature and precipitation estimates suggest this period was initially warm/wet turning to warm/dry.

Neoglacial pollen changes are evident at a variety of depositional sites in the northeastern Great Basin area. Depending on the sensitivity of the record, evidence ranges from a generalized cooler/wetter Neoglacial period to indications of vegetational changes associated with Neoglacial stades and interstades. Vegetational changes related to the earliest Neoglacial stade beginning about 5000 B.P. are evident at a number of sites (e.g., Baker 1976, Dalley 1977, Madsen and Currey 1979, Lindsay 1980). At least two subsequent periods indicative of cooler and/or wetter conditions separated by a warmer/dryer period are evident at Snowbird Bog, Crescent Springs (Mehringer 1977), Sudden Shelter (Lindsay 1980), and possibly at Swan Lake. These records appear to correlate, but the temporal controls at the sites are not sufficient to allow them to be tightly related, although the second cool/wet period appears to have ended by at least 2000 years ago at all locations. Vegetational change associated with a possible third Neoglacial stade is not evident at any of these sites, but the upper portion of most of these records has been disturbed.

Holocene Pollen Records From the Great Salt Lake

This rather cursory discussion of Holocene records from around the Great Salt Lake is intended primarily to illustrate the relative interpretive utility of pollen from the lake itself. Long-term vegetation changes, those older than the Late Wisconsin, will most certainly be determined primarily from pollen analysis of deeply bedded lake deposits rather than from plant macrofossil records and it is important to judge how well Holocene samples from the lake compare to similarly dated samples from the more sensitive surrounding locations.

One of the principal problems with the interpretation of samples from lake cores is that it represents vegetation from roughly 125,000 sq kms and pollen in the deposits is derived from both long- and short-range aerial transport as well as from water transportation in the rivers feeding the lake. Preliminary data from surface deposits around the lake floor show significant variation (for example, pine ranges from 7.5% to 25.9% in seven samples), depending on depositional environments and proximity to river mouths, and it is clear that additional information on the nature of pollen influx into the lake will be necessary for interpreting deeper cores (Madsen and Kay 1982). A second problem in comparing lake records with other nearby pollen sequences is the lack of chronological controls on cores from the lake. It is apparent that problems with redeposited carbonates hamper the use of ^{14}C dates from the lake and there are only a limited number of fixed dates derived principally from dated volcanic ash desposits (Spencer et al 1984).

There are a limited number of pollen sequences available from the Great Salt Lake. Paul Martin and Peter Mehringer (1965) analyzed selected samples from cores taken by A.J. Eardley from the Great Salt Lake Desert, but reported only a generalized pine/spruce curve for samples older than 25,000 years ago. Mehringer (1977) published a summary diagram of samples from a core taken at Crescent Spring, northwest of the lake, that span the last Pluvial and appear to extend into the transitional period, but did not publish data on the Holocene portion of the sequence. Pollen samples spanning the mid-to-early Pleistocene and extending into the upper Pliocene were analyzed as part of an AMOCO oil drilling operation, but the results have not been published (E.T. Peterson 1981, personnel communication). Mehringer, Nash and Fuller (1971) briefly allude to unpublished data on samples from a short core from near Bird Island that apparently spans the Holocene. Despite this apparent variety of records, little usable information has been published and the only record of some utility is that taken from a series of short-cores around the lake within the last several years (Spencer et al 1984; Madsen and Kay 1982).

The transition from Pluvial to post-Pluvial vegetation appears sharper in the lake samples than in those from surrounding deposits, but it appears that the geomorphology of the Bonneville Basin contributes more to this apparent disparity than does differential sensitivity to climatic change. At the end of the last Pluvial period, pollen of halophytic plants such as greasewood, saltbush and pickleweed appears suddenly in the record (virtually from one sample to the next) and dominates throughout the remainder of the sequence. Conifer pollen, which dominates through the Pluvial, does not show a comparably sudden drop, but rather gradually declines over a much longer time span in a fashion comparable to the non-lake records. This appears to be the result of a relatively vast expanse of flat-bottomed valleys becoming exposed abruptly with the fall of the lake to its modern average at the close of the last Pluvial. This change appears so suddenly in the record that closely bracketing radiocarbon dates should be able to date this event to within 1-200 years.

Change in the amount of pollen characteristic of salt-flat margin plants occurs twice throughout the remainder of the lake record. In both cases there is a reduction in Chenopods and similar plant types which suggest that substantial portions of the local valley bottoms were flooded. Lack of adequate dating hampers placement of these events, but the first occurred well prior to the deposition of Mazama ash about 7000 years ago and is probably related to the lake's rise to the Gilbert level about 10,500 B.P. (Currey 1980). The second event is even more poorly dated. Mehringer et al. (1971) correlate a change in Chenopodiaceae/ *Artemisia* ratios, a change from coarse to fine laminations and the absence of carbonate coated brine shimp feces with a similar change in pollen ratios at a more adequately dated lake-side bog, and suggest increased water depth in the Great Salt Lake about 2-3000 years ago. This decrease in relative amounts of Chenopodiaceae pollen is evident in a more recent series of cores at about the same stratigraphic position (Spencer et al 1984), although the presence of two separate periods of high water suggested by chemistry and other data could not be detected.

There is very little change in other plant types throughout the Holocene portion of the lake pollen record and without the supporting structural and geochemical evidence, even these two suggested periods of higher water would be difficult to detect and harder to support on the basis of pollen analysis alone. Periods of decreased effective moisture and reduced lake levels are even more difficult to detect and do not show up well in the Holocene pollen records. Again, this is due to a combination of geomorphology and the generalized nature of the lake pollen record. When the lake is reduced beyond its present size the proportion of salt-flat margin habitat is not greatly increased and, unlike during periods of higher water (greater than 1295m [4250ft]), there is little change in the ratio of halophytic plant types. Further, the pollen derived from long-distance transport is from such a wide variety of plant communities in such diverse settings that low-order changes are masked. In short, fluctuations on the order of Holocene changes are only detectable in the pollen record from the lake if they involve lake level increases sufficient to flood the Great Salt Lake Desert; vegetation changes associated with decreased lake levels are difficult to detect.

Late Wisconsin Pollen Records

Only two poorly dated pollen records spanning the last Pluvial are available from the Great Salt Lake. An incompletely published summary diagram from Crescent Spring spans a period from about 30,000 to 10,000 years ago (Mehringer 1977) and a second, more complete record, is available from a lake core and represents the last 30,000-plus years.

Although the two records differ in detail (probably as the result of different locations in the lake basin), they are in agreement in their essential points. During the last interpluvial, beginning prior to the beginning of the records and lasting until about 23,000 years ago, a number of features suggest a lake elevation of 4220-4250 ft. The sagebrush/conifer ratio is higher and is indicative of an interpluvial climate that was as dry, but possibly cooler than the Holocene average. Chenopodiaceae-Amaranthus pollen is higher than during the succeeding Pluvial, but much lower than during the Holocene, suggesting the salt-flat margin habitat was reduced, but not completely flooded. Crescent Spring, currently at an elevation of about 1295m (4250ft), was flooded during this period, but apparently only to a shallow depth since marsh vegetation was prevalant. It may be that the lake oscillated around this elevation, alternatively flooding and exposing the Crescent Spring area.

During the last Pluvial, both pollen sequences are almost totally dominated by conifers, with both Chenopodiaceae and sagebrush pollen substantially reduced. There has been some suggestion (Wells 1983) that at the height of the last Pluvial a coniferous forest extended virtually to the water's edge and that a limber pine forest covered much of the valley bottoms. Other plant macrofossil data suggest open valley floors (e.g., Thompson 1984) and a fauna consisting predominantly of grazers (Michael Nelson 1985, personnel communication) suggests a fairly extensive grass cover. This is consistent with the pollen data, since grass pollen remains relatively unchanged thoughout the length of both cores. Present data suggest a parkland, not unlike that in the Yellowstone area, with a mosaic of coniferous forest cover, probably on north- and east-facing slopes, and open sage/grass cover on south- and west-facing slopes.

There is some suggestion of a climatic oscillation during the last Pluvial in the pollen record from the lake core (Madsen and Kay 1982). Two periods of high relative and absolute amounts of conifer pollen are separated by an interval characterized by reduced pollen input and suggest a change in forest productivity and a corresponding climatic change. While it is interesting to speculate on a possible correlation with the proposed Keg Mountain oscillation (Oviatt 1984) which apparently occurred during this period, there is some suggestion of a depostional/ preservational change in the samples from this interval, and it is possible that the observed change is unrelated to climatic events.

Mid-Wisconsin and Earlier

Two limited records provide pollen data back to and into upper Pliocene deposits. The pine/spruce curve from the Great Salt Lake Desert cores (1965) appears to extend into the "Illinoian", while the AMOCO record (Earl Peterson 1981, personnel communication) extends past the Plio-Pleistocene boundary. The AMOCO record is not climatically very useful since each sample was "composited over each 100 feet (30m) of interval," but it does indicate that pollen is preserved thoughout most of the nearly 1000m of core length. Apparently "1400 feet (427m) of Pleistocene (deposits were) removed by erosion." The Great Salt Lake Desert cores contain no pollen in what is thought to be the Sangamon interglacial and in the underlying deposits, "what should be a glacial" is marked by pine/spruce ratios much below that in the Wisconsin. The Wisconsin record from the cores comes from samples underlying dates of "full-glacial age" and is marked by increases in pine/spruce ratios at the beginning and end of the record.

Determination of Long-term Climatic Fluctuations

Short-to-medium term climatic fluctuations in the Bonneville Basin will probably be determined primarily by a combination of plant macrofossil data and geophysical/geomorphological information, but pollen analysis will be an important component of long-term continental climatic reconstructions. Cores from Bonneville Basin deposits spanning 3 million years or more are available and contain relatively well preserved pollen throughout most of their length. Short and medium-length pollen records from in and around the Great Salt Lake provide useful interpretive analogs for these deeper deposits.

Several interpretive principles appear to be useful. Due to the shape of the basin, the relative amount of Chenopodiaceae is apparently a good guide to shallower water depths, while the relationship between conifer and sagebrush pollen appears to be useful for identifying deeper water conditions. In applying vegetative conditions to climatic conditions and in turn to lake levels, the following lake level conditions can probably be identified in deep lake cores (it should be kept in mind that these are equivalent water levels; non-climatically induced changes such as the capture of the Bear River must be examined independently):

1) Climatic conditions leading to high amounts of Chenopodiaceae-Amaranthus and sagebrush and low amounts of conifer pollen probably result in water level equivalencies similar to or lower than the historic average.
2) Climatic conditions leading to relatively low amounts of conifer pollen, moderate amounts of Chenopodiaceae-Amaranthus pollen, and relatively high amounts of sagebrush pollen probably result in water levels similar to that during the Gilbert or mid-Neoglacial.
3) Climatic conditions leading to high amounts of conifer pollen, low amounts of Chenopodiaceae pollen, and moderate amounts of sagebrush pollen probably result in water levels similar to Provo/Bonneville levels.

By using these and other interpretive techniques on a fine scale pollen sampling of deep-cores from the lake, and combining the results with other available geophysical information, the Pleistocene record from the Bonneville Basin can be made the continental equivalent of the marine record. When combined with long records from other western North American lakes, such as Searles Lake and Clear Lake, it should be possible to identify similarities and differences between marine an northern hemisphere continental climates for the whole of the Pleistocene.

Perhaps what is more important for estimating future levels of the Great Salt Lake is that detailed long records will allow the identification of long-term trends against which the extremely short historical climatic data can be evaluated. Should these long term trends be towards increasing or decreasing effective moisture, rather than the flat-base that meteorological and climatic models must currently assume, then the nature of lake level predictions must be significantly altered. To put it another way, the rapid rise of the Great Salt Lake in the early

1980s may not be unusual or above average, but rather may represent "normal" conditions if the over-all long-term trend is towards increasing precipitation. Clearly, the determination of long-term climatic trends is critical to the understanding and management of the Great Salt Lake.

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Tree-Ring Inferences on Historical Changes in the Level of Great Salt Lake

by

David M. Meko and Charles W. Stockton
Laboratory of Tree-Ring Research
University of Arizona
Tucson, Arizona
March 1985

Introduction

Following extreme rises in water level in the past two and a half years, the level of Great Salt Lake, Utah, now (spring, 1985) is higher than at any time since reliable record-keeping began in the 1870's (Arnow 1984; U.S. Geological Survey, unpublished graph). If past lake-level variations are to be used as a guide in developing models for predicting future variations, it is important to gain a long-term perspective on lake-level variability. Tree-rings can be a useful tool in quantitatively extending our knowledge of lake-level history on the time scale of years to centuries.

Tree rings have previously been used to infer variations in lake levels in the Great Basin by Antevs (1938) and several other researchers working in the first half of the twentieth century. Harding (1965) compared the results of several tree-ring studies of lake levels in the Great Basin and concluded that there was little evidence for climatic change over the past 500 years. Both Antevs' study and those reviewed by Harding were done on lakes in the western part of the Great Basin, however, and their results are probably not applicable to the Great Salt Lake. The importance of having adequate spatial coverage by tree-ring sites was emphasized in a more recent reconstruction of streamflow in the Upper Colorado River Basin (Stockton and Jacoby, 1976), in which large differences were found in tree-growth variations from northern and southern tributaries of the Green River. The drainage of the Green River is adjacent to that of Great Salt Lake.

The purpose of this paper is to examine previously collected tree-ring data for evidence of changes in the level of Great Salt Lake. The currently available tree-ring network is deficient in that no sites are within the drainage basins of the lake's tributaries. Sites collected for other studies over the past two decades do allow us, however, to examine the general climatic history of a rather large geographic area which includes Great Salt Lake and its tributaries.

Data

The tree-ring data were collected in various field expeditions by researchers from the Laboratory of Tree-Ring Research, University of Arizona (TRL), between the years 1964 and 1982. A total of 11 tree-ring sites were used; locations of the sites are marked on the map in Figure 1, and identifying information is given in Table 1. The 11 sites were screened from a considerably larger number of sites using the following criteria: 1) upper timberline sites were excluded 2) sample size (number of cores) was required to remain reasonably stable back in time to 1700 and 3) sites collected before the year 1964 were excluded. To retain sufficient sample size for a large-scale analysis only the years after 1700 were included in the analysis.

The tree-ring data were obtained from the TRL files in the form of indices[1] for each site. Autocorrelation was then removed from each site's tree-ring index by filtering the series with an appropriate low-order (third order or less) autoregressive model (Box and Jenkins,1976). The objective of this filtering or "whitening" is to compensate for the natural filtering of the climatic signal by the biological system of the tree

Table 1. Tree-Ring Sites

MAP[1] NO.	STATE	SITE[2] I.D.	SPECIES	ELEVATION (m)	LAST YEAR	AR ORDER
1	Utah	123549	Ponderosa pine	1920	1964	2
2	Utah	125649	Ponderosa pine	2621	1964	3
3	Utah	279540	Douglas fir	2286	1971	1
4	Utah	280620	Colorado pinyon pine	2286	1971	2
5	Utah	281550	Engelmann spruce	2865	1971	1
6	Wyoming	552590	Limber pine	2377	1971	2
7	Wyoming	282540	Douglas fir	2652	1971	1
8	Wyoming	283590	Limber pine	2499	1972	3
9	Nevada	WH5639	Singleleaf pinyon pine	2100	1982	1
10	Nevada	PEX639	Singleleaf pinyon pine	2420	1982	1
11	Idaho	CORLP9	Singleleaf pinyon pine	2130	1980	1

[1] map shown in Figure 1.
[2] identification code for site.
[3] year of collection of site.

[1]the tree-ring index X_t is defined as $X_t = W_t/G_t$, where W_t is the ring-width and G_t is the value of the fitted age curve in year t. Indices from individual cores from many trees are averaged to form what is called the "site" tree-ring index.

(Meko, 1981). The orders of autregressive models used are listed in Table 1. Finally, the 11 whitened tree-ring series were regionally averaged to represent tree-growth variations in four regions: east-central Utah, the Uinta Mountains of northeastern Utah, the Wind River Mountains in Wyoming, and northeastern Nevada.

The water-level data for Great Salt Lake were obtained from Paul Kay of the University of Utah. The data for 1851-1968, mean annual (water year) values adjusted for distortion due to human interference, originally came from the Utah State Division of Water Resources. The data for 1969-1984, observed first-of-month values, came from U.S. Geological Survey Water Supply Papers and Ted Arnow of the Survey. These latter data were made compatible with the 1851-1968 data by adding a figure of 5.44 feet--a human interference adjustment--uniformly to all monthly values and averaging to approximate water-year averages. The quality of the lake-level data is not equally good for all parts of the 1851-1984 record (Arnow, 1984).

Two points should be noted in attempting to establish a quantitative relationship between tree rings and water levels of Great Salt Lake. First, the areas contributing inflow to the lake are not sampled by the available tree-ring sites. Over 90 percent of the average annual inflow comes from either direct precipitation on the lake (31 percent) or surface flow from the Bear, Weber, and Jordan Rivers (about 60 percent) (Arnow, 1984). Only one tree-ring site, on the north side of the Uinta Mountains, is near a drainage basin of one of the tributaries. Second, because of the highly variable surface area of the lake, the relationship between lake level and variations in total runoff volume from the drainage of the lake is not even approximately linear.

Observed Lake-Level Changes

Since interannual climatic variations are more likely to be related to changes in lake level than to lake level itself, the first-differenced series of lake levels was selected as a variable for comparison with annual tree-ring series. Both the lake levels and first-differenced lake levels are plotted in Figure 2. The two series are dissimilar in many ways. Major peaks and troughs in the differenced series do not in general coincide with those in the original series. Differencing also clearly changes the relative magnitude of various peaks and troughs. For example, the differenced series emphasizes the wet period in the early 1900's relative to the period in the late 1860's and early 1870's. Major wet periods in the differenced series are indicated centered on 1868, 1908, and 1972; major dry periods on 1901, 1934 and 1961. The unprecedented jump in lake level over the past two years is also clear from Figure 2; but several more year's data are necessary to determine whether the most recent rises are part of a major low-frequency fluctuation like that of the early 1900's.

Regional Tree-ring Variations

Smoothed time series plots of the four regionally averaged tree-ring series (Figure 3) show that similar low-frequency variations in tree-growth have occurred at widely separate locations to the north, south, east and west of Great Salt Lake. The parallel behavior of the tree-ring series suggests that large-scale climatic variations over the western United States were a major determining factor in the tree-ring variations. Inter-regional correlation was highest between Northeastern Nevada and the Wind River Mountains (r = .65 for period 1703-1964). The lowest correlation was between the Wind River Mountains and East Central Utah (r = .31). There is unanimous agreement among the series that the period from 1905-1920 was the most favorable extended period for tree growth since 1700. The period with large increases in lake level in the late 1860's (Figure 2) also was characterized by greater than normal tree growth in all regions, but the anomalies were generally less severe than in 1905-20.

Imbedded in the generally parallel behavior of the regional tree-ring series are some notable spatial contrasts. For example, droughts centered at 1880 and 1902 might be classified as "southern" droughts; while droughts centered in the 1860's and near 1961 were "northern" droughts. Similarly, wet periods at 1726 and 1839 were markedly wetter to the north, the wet period at 1940 to the south.

Strength of Relationship between Lake Level Changes and Tree Rings

The similarity in many major time series features of the plots of lake-level changes and tree-ring variations (Figures 2 and 3) suggests that some quantitative lake-level information can be extracted from the tree-ring data. A correlation matrix for the lake-level and tree-ring variables for the period 1852 to 1964 is shown in Table 2. If significance is attached to coefficients whose magnitude is greater than $2/\sqrt{n}$, where n is the sample size in years, all four regional tree-ring series are significantly correlated with the first-differenced lake-level series. A fifth

Table 2. Correlation Matrix of Tree-ring[1] and Lake-level Variables 1852-1964.

		Tree-Ring Series				
	Lake Level	**E. Central Utah**	**Uinta Mtns**	**Wind River Mtns**	**N.E. Nevada**	**Mean Tree-Ring**
Lake Level	1.00	.32	.41	.36	.28	.42
E.Central Utah	.32	1.00	.62	.31	.42	.76
Uinta Mtns	.42	.62	1.00	.59	.60	.87
Wind River Mtns	.36	.31	.59	1.00	.65	.75
N.E. Nevada	.28	.42	.60	.65	1.00	.83
Mean Tree-Ring	.42	.76	.87	.75	.83	1.00

[1] see text for description of variables

tree-ring variable, the arithmetic mean of the four regional tree-ring series, is more highly correlated with lake-level changes than are any of the individual regional tree-ring series. The strength of the correlation ($r = .42$) is still relatively weak compared to correlations typically found in other tree ring-climate studies (e.g., Stockton and Meko, 1983; Blasing and Duvick, 1981); a reconstruction based on simple linear regression would only explain $(.42)^2 \times \cdot 100 = 18$ percent of the variance of lake-level changes.

Examination of lagged cross-correlations (not shown) between the lake level series and the mean tree-ring series revealed that information on the lake-level change in year t may be available from values of the tree-ring series in years t-1 and t-2 as well as in year t. The following multiple regression equation was consequently adopted for reconstructing lake-level changes:

$$\hat{Y}_t = \hat{a} + \hat{b}_0(X_t) + \hat{b}_1(X_{t-1}) + \hat{b}_2(X_{t-2})$$

where $\hat{Y}_t$ is the predicted lake level first difference in feet in year t,
X_t is the mean of the four regional tree-ring series in year t, and
$\hat{a}$ is the constant in the regression equation, and
$\hat{b}_0$, $\hat{b}_1$, and $\hat{b}_2$ are the estimated regression coefficients.

The estimated regression constant and coefficients were:

$$\hat{a} = -0.08898;\ \hat{b}_0 = +1.78936;\ \hat{b}_1 = +1.41363;\ \hat{b}_2 = +0.53648$$

The multiple correlation coefficient for the equation was $R = .54$, which corresponds to 29 percent variance explained. This is a considerable improvement over the variance expalined by a model without lags, but still is relatively poor compared to statistics tyically encountered in tree-ring reconstructions of precipitation or drought indices.

The estimated coherency-squared function (C^2) from a cross-spectral analysis (Jenkins and Watts, 1968) between the actual and reconstructed lake-level changes 1852-1964 (Figure 4) shows that the signal in the reconstruction drops off sharply at wavelengths shorter than about four years. The reconstruction is therefore most reliable in portraying lake-level changes smoothed over several years. Smoothed time series of the reconstructed and actual lake-level changes are plotted in Figure 5 for the 1852-1964 calibration period. The reconstructed and actual data agree fairly well on the timing of major droughts. The most negative smoothed values of lake-level change occurred at about 1934, 1960, 1880, and 1902 in both the actual data and the reconstruction. The reconstruction is considerably less accurate in pinpointing positive peaks in lake-level change, and certainly in establishing the relative magnitudes of these peaks. The reconstruction unduly emphasizes the wetness of the early 1900's relative to that of the late 1860's-early 1870's, and fails entirely to identify a peak in the actual data centered at 1951.

There are several possible reasons for the poor perfomance of the reconstruction in detecting the timing and magnitude of periods of very large increases in lake level. Negative precipitation anomalies are known to be generally more spatially coherent than positive anomalies (Julian, 1970). The inadequate tree-ring sampling of the areas contributing runoff to Great Salt Lake is therefore more likely to lead to errors in detecting wet peaks than drought. Also, if the lake-level rise is due to runoff from only a few storms with extremely intense rainfall--perhaps from heavy thunderstorms in the warmer months of the year--tree rings may not adequately reflect the magnitude of the wetness; the sensitivity of tree growth to rainfall variations may decrease drastically beyond some optimal level of soil moisture. Finally, the quality of the lake-level data itself falls off before the year 1875: the magnitude of the observed lake-level anomaly in the late 1860's and early 1870's may not be as severe as the observed data implies.

Reconstructed Lake-level Changes Earlier than 1852

Time series plots of the smoothed changes in reconstructed lake level (Figure 6) show that the dry period centered at 1934 and the wet period at 1914 were unequaled in severity at least back to A.D. 1709. The occurrence of the second wettest reconstructed value in 1908 attests to the extreme general wetness from about 1905 to 1920. A similar result was reported by Stockton and Jacoby (1976) for the reconstructed annual virgin streamflow at Lee Ferry, Arizona. This is no coincidence since both studies used many of the same tree-ring data. Reference to the tree-ring plot for the northeastern Nevada region (Figure 3) shows, however, that the anomalous wetness of the 1905-1920 period extended well outside the drainage of the Colorado river. The five periods of greatest reconstructed lake-level increase and decrease were centered on the following years:

INCREASE	DECREASE
1914	1934
1908	1709
1726	1824
1839	1782
1868	1846

Results so far suggest that the dates listed above for decreases in lake level are far more reliable than the dates for increases.

In spite of the noted difficulties with reconstructing rises in lake level, it is interesting to place the observed 1983-84 rise (Figure 2) at least roughly in a long-term context. The observed 4.4 ft. rise in 1983-84 compares with extremes of reconstructed rises of 1.0 ft. in 1907-08 (post-1851 extreme) and 0.9 ft. in 1726-27 (pre-1851 extreme). (Note that only the smoothed reconstruction is plotted in Figure 6.) The standard error of the reconstruction was 0.74 ft. Both the 1907-08 and 1726-27 reconstructed rises are more than 4 standard errors below the observed 1983-84 rise. It is therefore unlikely that the 1983-84 rise has been exceeded in any year at least back to 1703.

Conclusions

1. Changes in water level of Great Salt Lake are significantly correlated (.05 confidence level) with regionally averaged tree-ring series from areas surrounding the lake, but outside its drainage. Large-scale climatic fluctuations are probably the cause of this correlation.

2. The lake-level signal in tree-ring series is concentrated at frequencies lower than about (1/4) yr^{-1}.

3. The available tree-ring data indicate that the wettest climatic conditions since 1700 in a large geographical area surrounding the lake occurred in the first two decades of the twentieth century. The driest conditions were in the 1930's. Tree-rings record two periods at 1908 and 1914 as wetter than the late 1860's; this result appears to contradict the observed lake-level data.

4. The observed single-year rise in lake level from 1983 to 1984 is probably unequaled at least back to 1703.

5. The signal for lake-level changes in the available tree-ring data is probably too weak to be of much use as supplementary information in models for predicting future variations in level of Great Salt Lake. Additional tree-ring collection could greatly alter this conclusion. A considerable gain in the strength of the signal would be expected if a tree-ring network was collected from the Wasatch Range and other areas contributing inflow to the lake. The time period of reconstruction could probably be extended back to A.D. 1600 with a new collection, and the ability of the tree-ring data to reconstruct huge lake-level rises like that of 1983-1984 could be checked.

Acknowledgements

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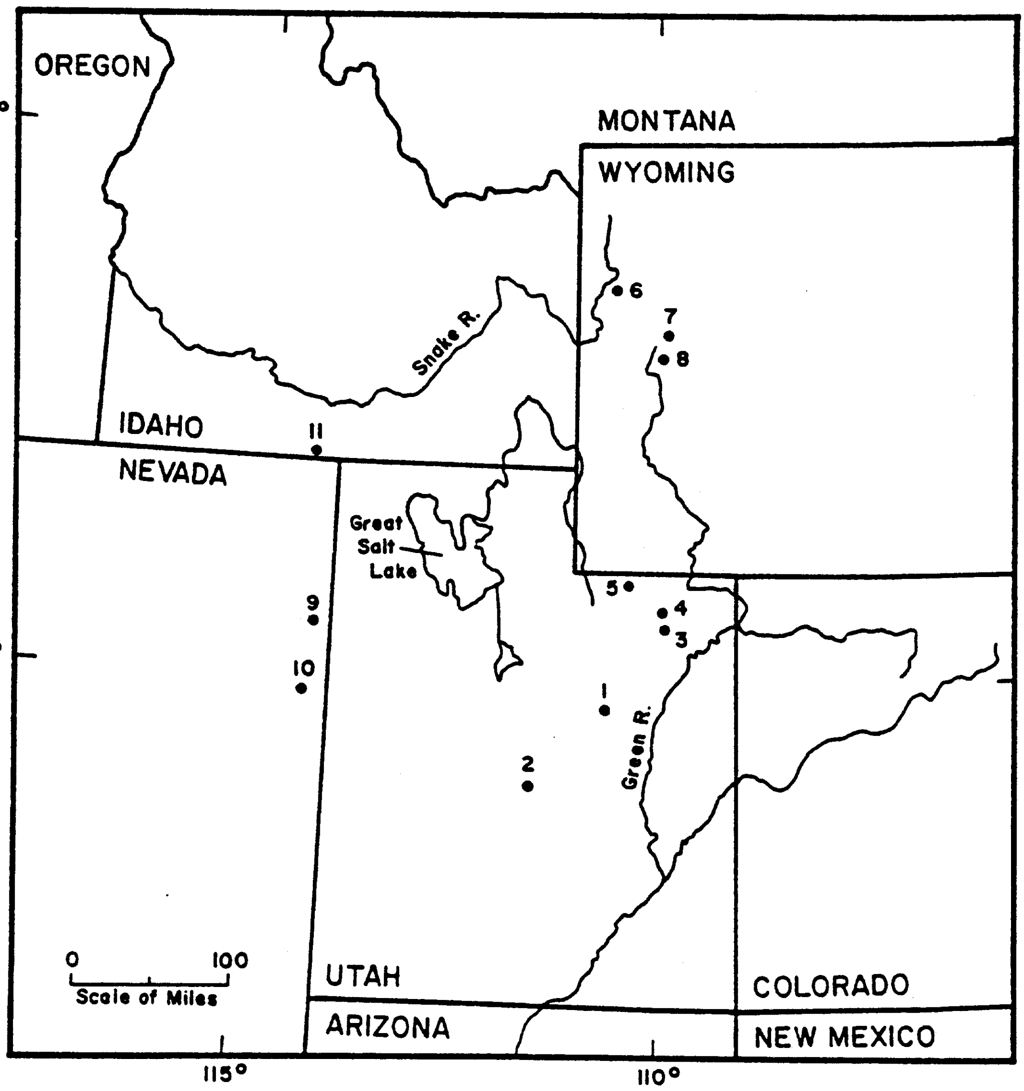

Figure 1. Map showing locations of tree-ring sites. Numbering follows Table 1. Sites were grouped into four regional series: E. Central Utah (sites 1,2); Uinta Mtns (sites 3,4,5); Wind River Mountains (sites 6,7,8); and Northeastern Nevada (sites 9,10,11).

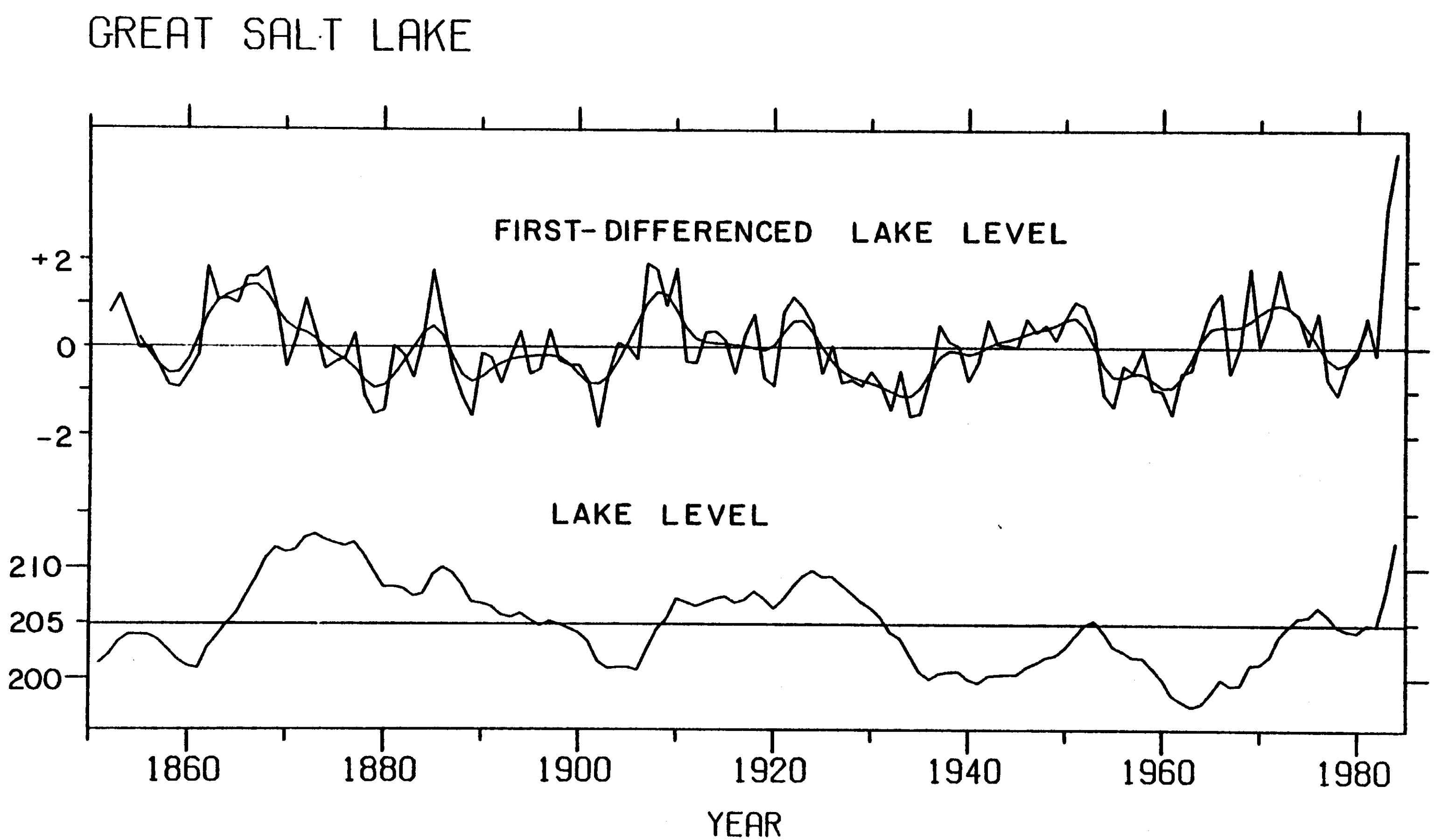

Figure 2. Time series of observed level of Great Salt Lake. Bottom: mean water-year average lake level above a datum of 4000 feet for period 1851-1984. Top: first-differenced lake level. The first-differenced lake level is a time series of year-to-year changes in lake level. A low-pass-filtered plot of the lake-level changes is also shown. The filter was a raised cosine (Hamming, 1977) with weights .2646, .2187, .1158, .0332 on lags 0 through +3 years. This filter virtually eliminates variance at wavelengths shorter than 4 years. The values of the frequency response at wavelengths 20, 10, 5 and 4 years are 0.91, 0.67, 0.16 and 0.03. Note the different scales for the top and bottom plots.

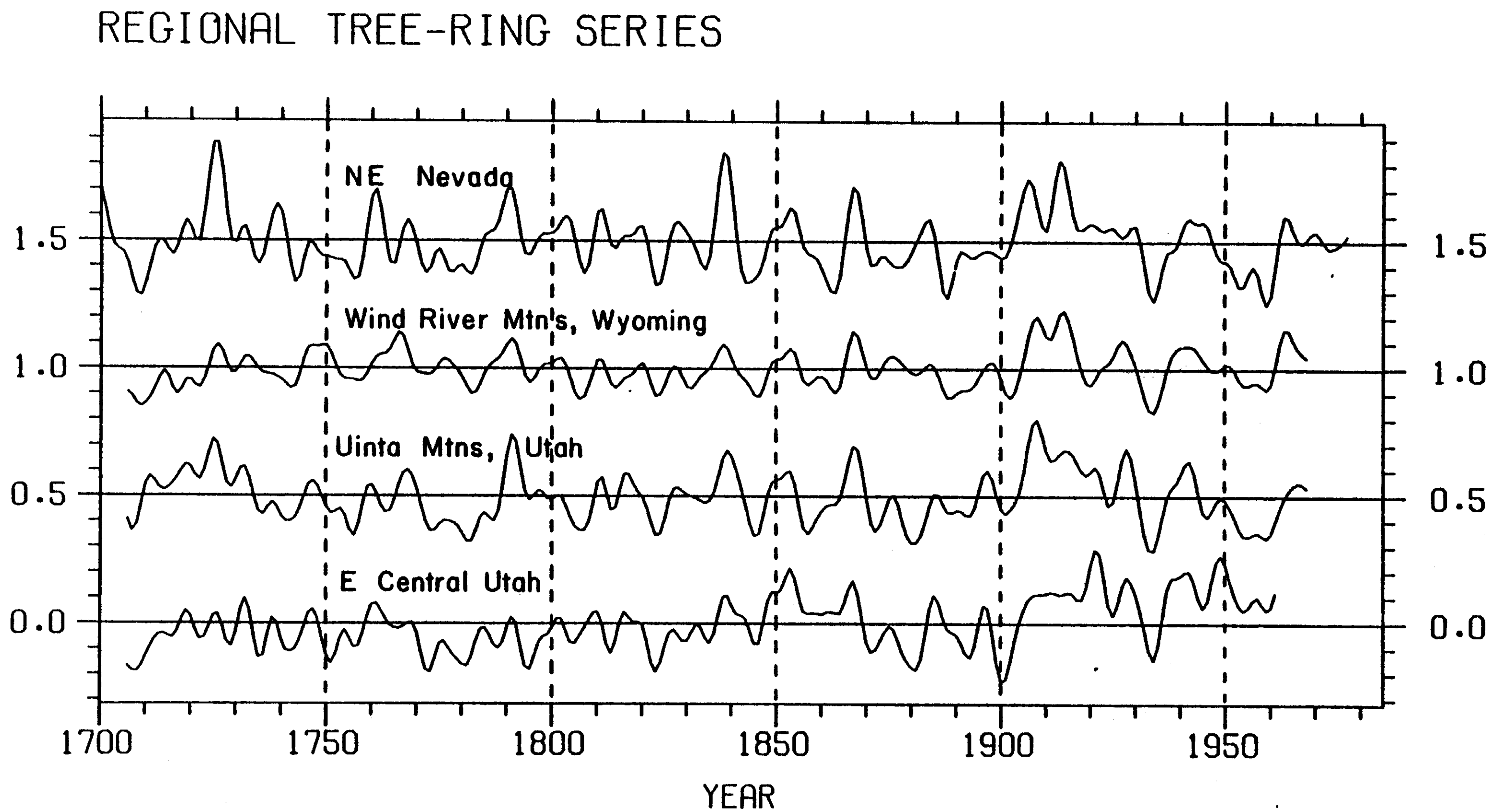

Figure 3. Filtered time series of regional tree growth. Series plotted are regional averages of "whitened" tree-ring indices smoothed by the same low-pass filter described in the legend for Figure 2. Increments of 0.5 have been added to the top three series before plotting. A deviation of x from the mean-line corresponds approximately to (1 + x) 100 percentage of normal tree-ring widths. For example, a value of +0.3 for the bottom plot corresponds to 130 percent of normal growth.

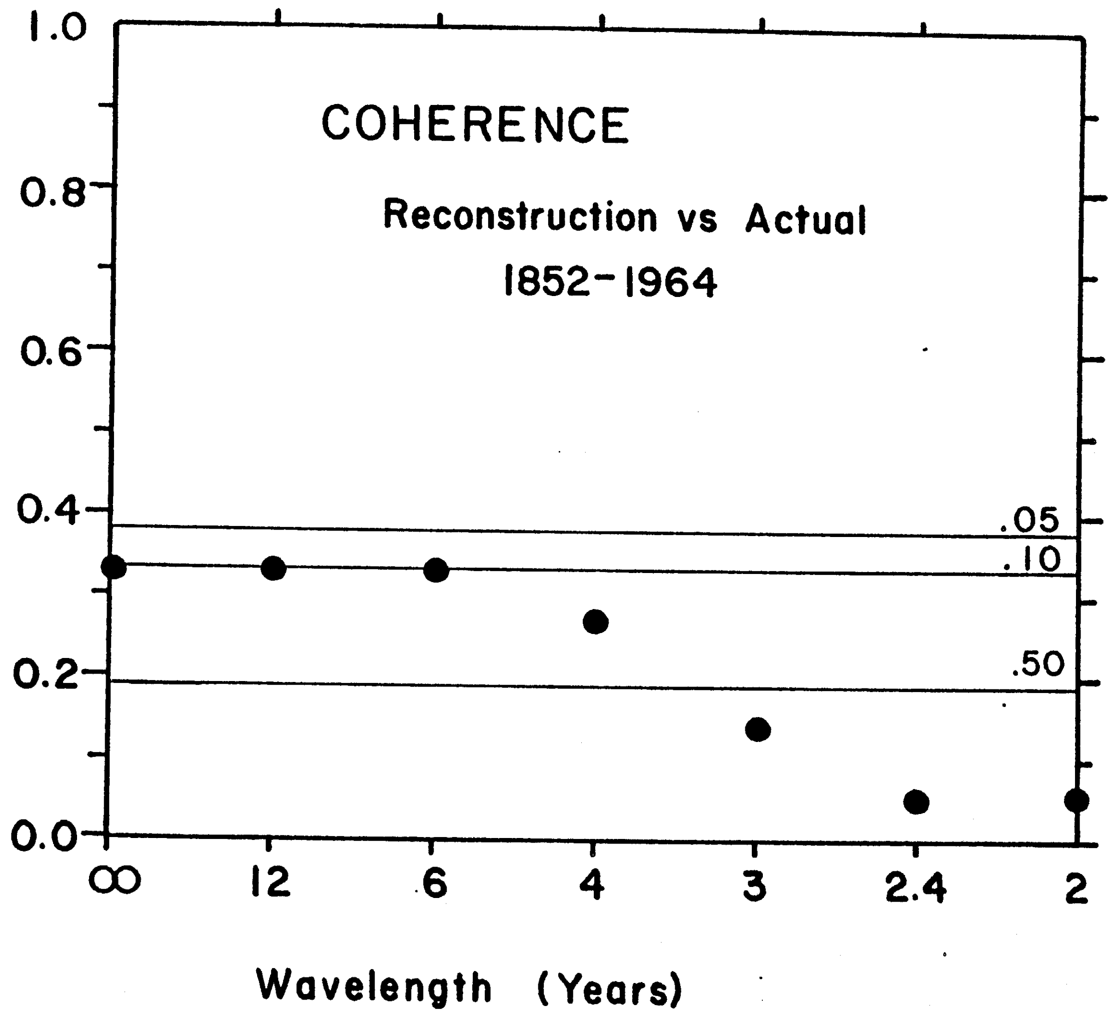

Figure 4. Estimated coherence between reconstructed and actual lake-level changes 1852-1964. The Blackman-Tukey method with a Parzen window with 6 lags (Jenkins and Watts, 1968) was used to estimate the cross-spectrum. The .05, .10, and .50 significance levels following Granger and Hatanaka (1964) are also shown.

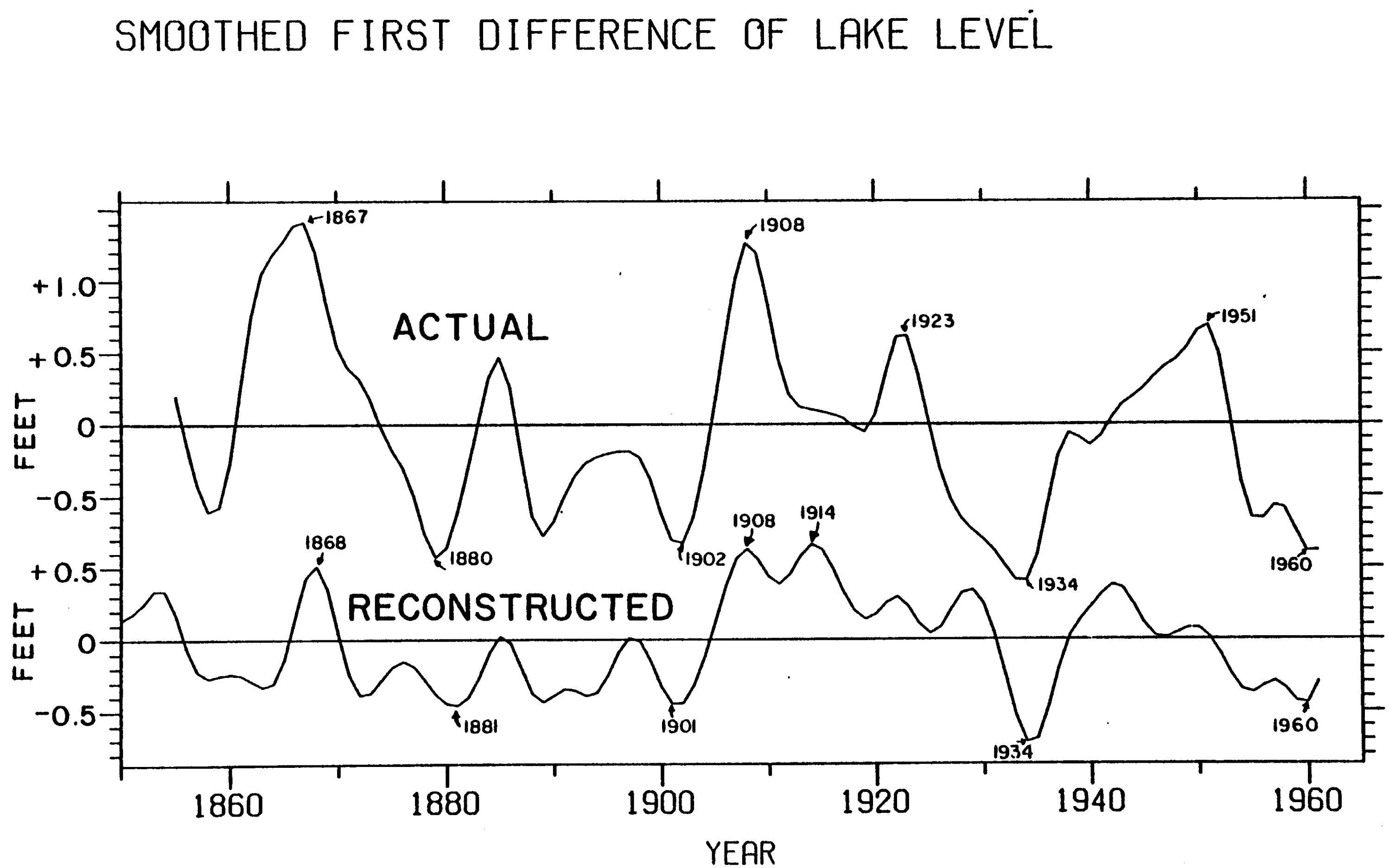

Figure 5. Time series plots of actual and reconstructed lake-level changes, 1855-1961. Data are first differences of lake level smoothed by the same raised-cosine filter described in the legend for Figure 2. Years of largest peaks and troughs are marked.

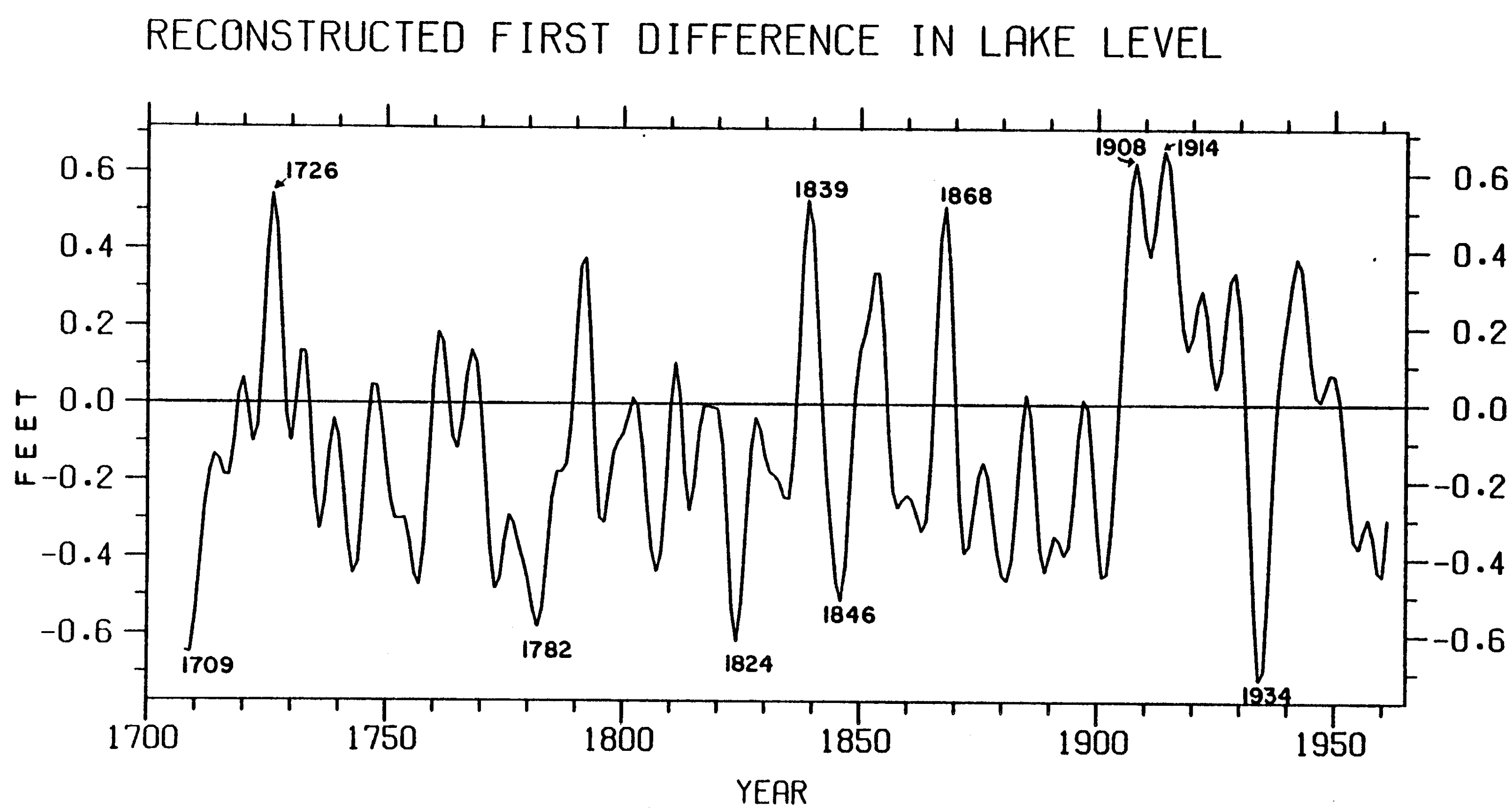

Figure 6. Time series plot of reconstructed lake-level changes to A.D. 1709. A low-pass filter (see legend for Figure 2) was applied to the annual reconstructed values before plotting. Years of major peaks and troughs are marked.

SESSION 2

HISTORICAL RECORD OF MEDIUM- AND SHORT-TERM FLUCTUATIONS

Water-Level Changes in Great Salt Lake, 1843-1984

by Ted Arnow, U. S. Geological Survey,
Salt Lake City, Utah

The surface of Great Salt Lake was 4,200 feet above sea level in 1843 according to a determination by John C. Fremont, and it was about the same level in 1847 when the Mormon pioneers arrived in Utah. At this level, the surface area of the lake is about 1,700 square miles. Continuous records of lake-level changes based on actual measurements or carefully calculated estimates extend from 1847 to the present. (See accompanying graph.) The lake rose almost 5 feet by 1855 but then declined again to 4,200 feet by 1860. From 1862 to 1873, the lake rose almost 12 feet to reach a historic high of about 4,211.6 feet. At this level, the surface area of the lake was about 2,400 square miles.

During the next 31 years, the lake declined almost 16 feet, and by 1905 it was at a then historic low of slightly less than 4,196 feet. A similar overall decline followed during the next 58 years, and by 1963 the lake had reached an all-time historic low level of 4,191.35 feet. At this level, the surface area was only about 1,000 square miles. The fluctuations of the lake generally reflected similar trends of precipitation as represented by the record for Salt Lake City where systematic record-keeping of precipitation was started during 1874.

In 1964, the lake began to rise in response to increased precipitation, and by 1976 it had risen 11 feet to slightly above 4,202 feet. But it began to decline in 1977 in response to unusually meager snowfall during the winter of 1976-77. Early in September 1982, the lake was at about 4,200 feet--the same level that it was 139 years earlier when visited by Fremont.

On September 18, 1982, the lake began to rise in response to a series of storms that occurred earlier in the month. The precipitation of 7.04 inches for the month (compared to an annual average for 1875-1982 of about 15 inches) made it the wettest September on record for Salt Lake City. The lake continued to rise for the next 9 1/2 months as a result of greater than average precipitation during the autumn of 1982, greater than average snowfall during the winter and spring of 1982-83, and unseasonably cool weather during the spring of 1983. The rise from September 18, 1982, to June 30, 1983, was 5.1 feet, the greatest seasonal rise ever recorded.

The lake declined only 0.5 foot during the summer of 1983 due to greater than average precipitation and less than average evaporation. The excessive precipitation continued throughout the autumn and culminated in the wettest December ever recorded at Salt Lake City. During 1983, the Salt Lake City International Airport received 24.26 inches of precipitation, about 1.6 times the average.

The cumulative precipitation from January to June 1984 also was greater than average, and the lake rose 5.0 feet from September 25, 1983, to July 1, 1984. This was the second largest seasonal rise ever recorded for the lake. The net rise of the lake from September 18, 1982, to July 1, 1984, was 9.6 feet. By comparison, the previously recorded maximum net rise in a 2-year period was 4.75 feet during 1970-72.

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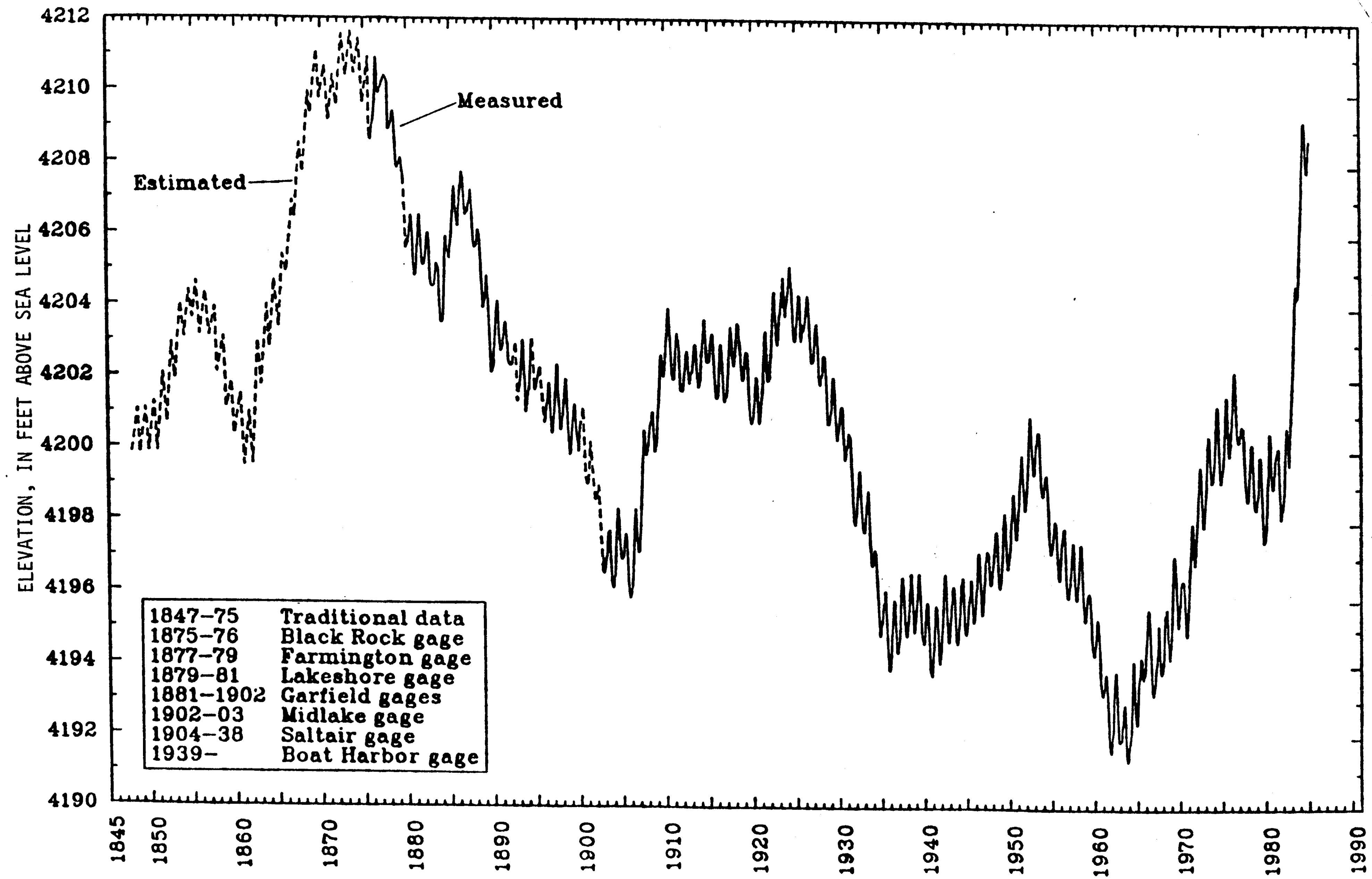
Fluctuations of the level of Great Salt Lake



HISTORICAL CLIMATE DATA OF THE GREAT SALT LAKE WATERSHED: Relation to lake levels

G. E. Bingham, E. A. Richardson and G. A. Ashcroft
Office of the Utah State Climatologist, UMC - 48
UTAH STATE UNIVERSITY, Logan, UT 84322

INTRODUCTION

Levels of the Great Salt Lake (GSL) have concerned the people of the Salt Lake region since the settling of the area by Mormon pioneers in the mid-1800's. Initially, concern was with the rising lake. After the lake peaked in 1877, the nearly 100- year downward trend of the lake level led many to conclude that the lake might at some future time completely disappear. Many of the studies conducted only twenty years ago, when the lake was at its all time low, were concerned with methods of preserving the lake. Since that time, the lake level has been on a nearly continuous increase; which linked with the nearly 8-foot rise in the lake level in 1983 and 1984, has returned the lake to near record levels. Because the lake is in a shallow basin, even small changes in lake level lead to large changes in its shore line. Such a variable shore line makes commercial development of the lake extremely difficult. After such a long period of decline, the relatively quick rise of the lake resulted in damage to surronding property estimated at nearly $200 million.

This paper examines the historic data collected by instrumentation since the settling of the region and its relationship the GSL levels. It also examines the Tree Ring Index for a few sites around the GSL basin to determine if sufficient correlation exists between ring width changes at these sites and lake levels or precipitation patterns to allow us to extend the record beyond the 100 to 130 years for which we have reliable recorded data.

HISTORIC CLIMATE DATA

On February 2, 1870 the following message was transmitted by the Signal Corp over the telegraph from Corrine, Utah Territory:

> Corrine, Utah Territory--Second day of month, A.M. observation: pressure reading not taken: temperature, 20 degrees F; humidity reading not taken; clear with wind from northwest; wind velocity not taken; no high clouds; no low clouds; and no precipitation past 12 hours. (quoted from Barfus *et al.*, 1967)

This was the first official weather observation from the State of Utah. Precipitation records from Corinne are continuous from that reading to present, but a gap exists in the temperature record from 1874 to 1895. Corrine was reopened as a climate station in January of 1897 and has the advantage that all observations taken at the station were reported from sites within a few blocks of the original location.

Interest in the weather of the area predates this event by several decades. Many of the early pioneers recorded specific weather events in their journals or diaries, and as early as March 6, 1852 the Deseret News made the following plea to its readers:

> "Temperature Records -- Will our friends who have thermometers in the various settlements of the mountains commence a daily reading on the 21st of April and report a copy of their readings every threemonths to the librarian of the Utah Library, Mr. William C. Staines. This will add one long-sought item to scientific history, which the world has hitherto sought for in vain."
>
> "Explanation: Place your thermometer on the north side of your house or building, about 10 feet from the ground; protect it from winds and storms, and at 6 in the morning, and 2 and 9 in the afternoon each day , note in your table the point where the mercury stands at those hours." (Quoted from Barfus et al., 1967)

The response to this request was not the best, since only a few sparse reports were ever published. On January 7, 1857 the Deseret News reported that, at the request of Governor Brigham Young, W. W. Phelps with the assistance of his son, Henry, had consented to keep a regular set of meteorological readings. Some of these records are still available. M. E. Jones collected some of the Phelps observations recorded during the period from 1857 to the early 70"s from the daily information published in the Deseret News, edited them and summarized them into monthly tabulations. The same method of observing temperatures continued until the late 1880's. Average temperatures published in these early records were average values of the three observations each day. When maximum and minimum thermometers became available, the average values were obtained by adding the maximum and minimum temperature values and dividing by 2.

In 1874 the official weather station was moved from Corinne to Salt Lake City where observations continued until 1954, at which time the City Office station was closed. A significant discontinuity exists between the Phelps record as published by Jones, and the more recent City Office record. In addition, the Salt Lake City Office station was moved several times during its existence. Occasionally, the station was at ground levels, but more often it was located on roof tops through the downtown area. Hence, the records are not as compatable as one would like. In 1928, a weather station was opened at the Salt Lake City Airport and as weather recording activities became more aviation oriented, this station came to be accepted as the official station for the area (see Figgins, 1984 for details).

In the 1880's, increased interest in meteorological observations is evidenced by an increased number of weather stations under the sponsorship of the Smithsonian Institute. Then, after several years of effort, a bill authorizing establishment of the Civilian Weather Bureau under the Department of Agriculture was signed by President Benjamin Harrison on October 1, 1890. This agency took over both the observational and forecast programs of the Signal Corp. The Weather Bureau added many new climate stations to the western portion of the United States. The beginning of World War 1 in Europe and the demand for increased food production stimulated the initiation of additional climate stations in the west. Since the early 1900s, fairly consistant data is available for the populated regions around the GSL. The existing NOAA cooperative data collection network is shown in Fig. 1. Most

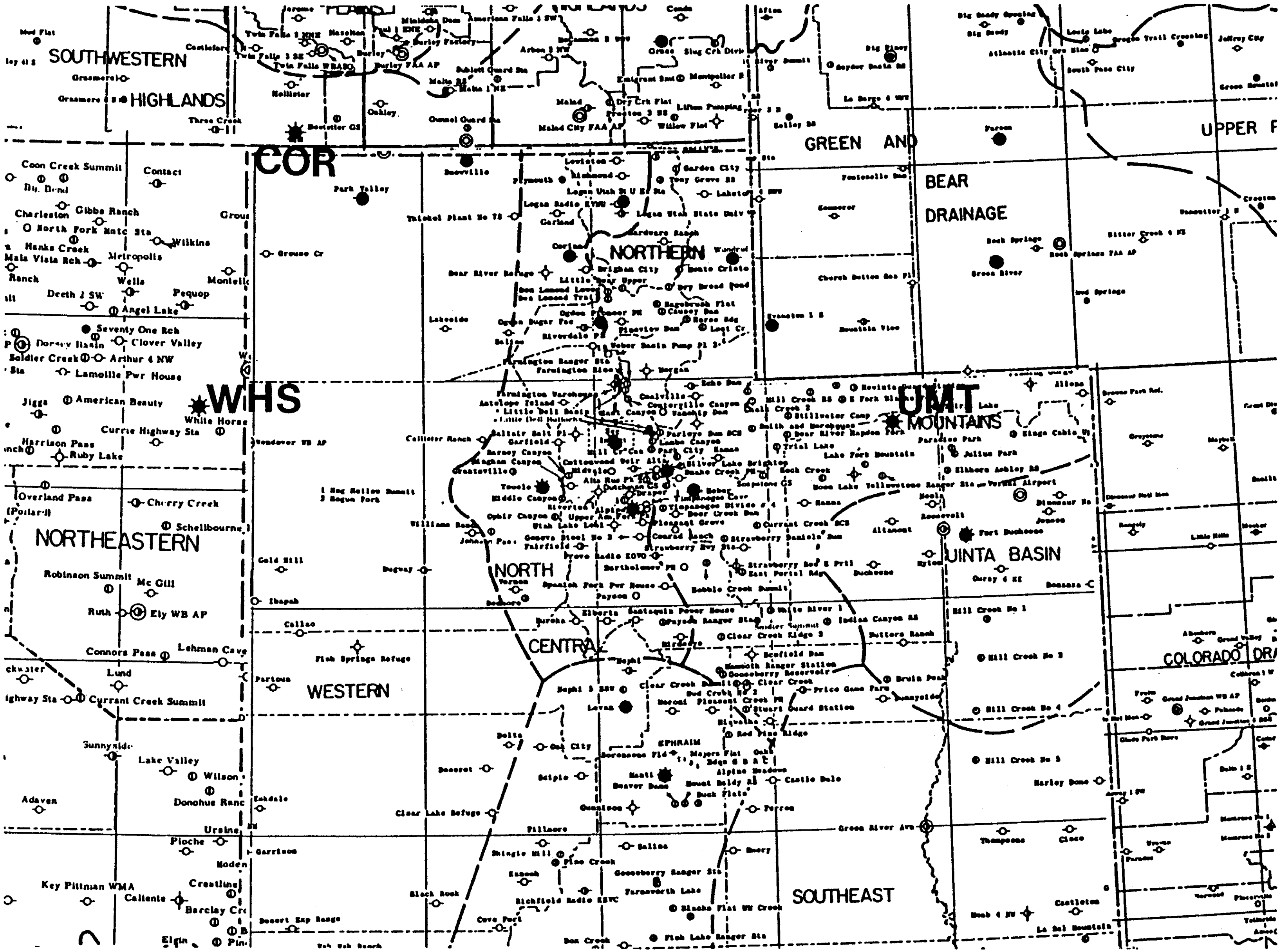


Fig. 1. A composite map of the NOAA Cooperative Network climate measurement stations around the Great Salt Lake Basin. Stations with periods of record which begin before 1905 are circled. The locations of the three tree ring sampling sites whose data are used in this report are also shown. These locations are starred and labeled.

of these stations collect precipitation and temperature data on a daily basis. Those stations having a period of record dating to the turn of the century have been highlighted in the figure. In recent years, some of the early period stations have been lost, due to funding reductions and a lack of concern for climatological information.

RELATION BETWEEN HISTORIC PRECIPITATION AND LAKE VOLUME

Before retiring as State Climatologist, E. A. Richardson made several studies comparing various averages of precipitation and GSL levels. Most of the prominant stations have serious continuity problems. The Salt Lake City record is an example. The City Office record is one of the earliest in the state, but only continues until 1954. During that period, the station was moved 6 times, some of which seriously effect the data. After 1954 airport data must be used. These differ significantly from the downtown record. In the comparisons used in this section of the report, we chose to use the City Office record, supplimented by regression generated data after 1954. The regression between the City Office and Airport precipitation was found to be:

$$\text{Precip (CO)} = 2.62299 + 0.9453 \times \text{Precip (AP)}$$

and was developed using data collected during the 26 years that both stations were in operation.

The recorded hydrograph of lake level (see Arno, This volume) is not an ideal record for comparison with climatic forcing functions. Not only is the early part of the record subject to estimation errors, but various amounts of water have been withdrawn from the system for irrigation and reservoir filling over the years. To help account for these changes in water use (see Stauffer, this volume) the amounts of water used has been estimated and added to lake volume allowing the construction of adjusted lake stage traces. Two adjusted hydrographs were prepared, one simulating the lake as though the river systems had always been fully developed (Modified), and one that an estimated lake stage as if man had never entered the system (Pristine). Because the relationship between volume and stage is not linear, a relationship to lake volume shows a significantly improved relationship to climatic factors. The indicator of lake state used in most of the comparisons reported here use Pristine Lake Volume as a dependent variable. Some of the early relations between precipitation and lake state use actual lake stage, but these were choose for time periods of relatively uniform water use.

After many trial comparisons, Richardson concluded that an 11 year running mean best correlated with lake level and volume changes. We do not wish to enter into a discussion of cause for this relationship, and for those who object to the use of this particular number, may substitue 12 or 13 for the rest of the paper without significantly altering the conclusions. The correlation between averaging time and lake level is detailed in Table 1.

It can be seen that a rather stable relationship exists for running means of annual precipitation with averaging periods of from 10 to 13 years. A similiar relationship was also found during other uniform withdrawal periods. Such a relationship was also found to exist with the individual station data, but the relationship was strengthened by the use of both stations. The addition of Heber to the relationship did not significantly improve the predictive power of the two station regression. At first it was assumed that the increase in predictive power from including Corrine was due to lake effect precipitation at SLC at higher lake levels. After completing the tree analysis in this paper, we conclude that shifting storm track patterns also effect the relationship.

Table 1. R^2 values for multiple regressions of various annual running mean averages of Salt Lake City Office and Corinne precipitation data against Great Salt Lake Level during the period 1956 to 1982.

Averaging period	1	2	6	8	10	11	12	13	14
Adjusted R^2	0.12	0.44	0.80	0.89	0.92	0.92	0.92	0.91	0.86

Over the whole period of the historic record, the regression model between precipitation near the GSL and pristine lake volume was not very good. Multiple regressions between 11-year running means of precipitation at SLC and Corrine and pristine volume show an adjusted R^2 over the 1885 to 1984 period of 0.43 and a multiple R of 0.66. The model seriously underestimates the extremes of the dependent variable. Over specific 30 to 50 year time periods, however, the predictive power of the model approached 90%. When the model was applied as a preditor of a comming years lake stage, by including the previous years lake volume as a dependent variable, an adjusted R^2 of 0.96 was achieved.

The poor correlation between precipitation and lake stage is not surprising when one considers that the collection basin for the lake is significantly larger than the region represented by the two stations used in the model. The significance of this effect is shown in the analysis of a more complete data set presented by Karl and Young, in this volume. An additional cause for the low predictive power of the simple regression model lies in the poor relation which often exists between valley and mountain precipitation. Temperature effects on evaporation and snow melt rate also modify the water harvest and are not reflected in the precipitation data from these two valley stations.

TREE RING INDEX AND LAKE VOLUME

Since the historic precipitation, temperature and lake stage records are so short, we examined some of the tree ring index data for the region in hopes that this data could be used to indicate the representativeness of the historic lake and climate data. The tree ring index data was provided to us by the University of Arizona Tree Ring Laboratory (Stockton, 1985, Personal communication, see Stockton and Meko, this volume). The data set consisted of normalized ring width data for 11 sites near the GSL basin, but none within the basin itself. Nine of the sites were in Nevada, with one in Idaho (City of Rocks) and one from near the center of the Uinta Mountains. After an initial prescreening, one of the Nevada sites (White Horse Summit - WHS), the Utah site (UMT) and the Idaho site (COR) were chosen for detailed comparison with pristine lake volume. Contiguous data sets for the four variables were available for the period 1851 to 1971. To be consistent with the precipitation analysis, 11-year running means were used, with the tree ring data beginning in 1841 so that a data point was available for each year lake volume was available. COR and WHS data were available until 1982, but the

UMT record terminated in 1971. Plots of the averaged tree ring data are shown in Fig. 2.

Like the precipitation data, the straight multiple regression of the averaged index from the three sites (see Fig. 3) with lake volume was poor (R^2 = 0.36). However, a close examination of Fig. 2 shows much better correlation between some of the sites and the lake during specific periods. For instance, the lake peak and decline between 1865 to 1905 (Fig. 4) showed a R^2 of 0.83 relationship to the averaged COR data. Between 1895 and 1971, (Fig. 5) the two more southerly tree ring sites, WHS and UMT, showed a adjusted R^2 = 0.68. This difference in predictive power leads us to conclude that a significant shift in the nature of the precipitation pattern of the

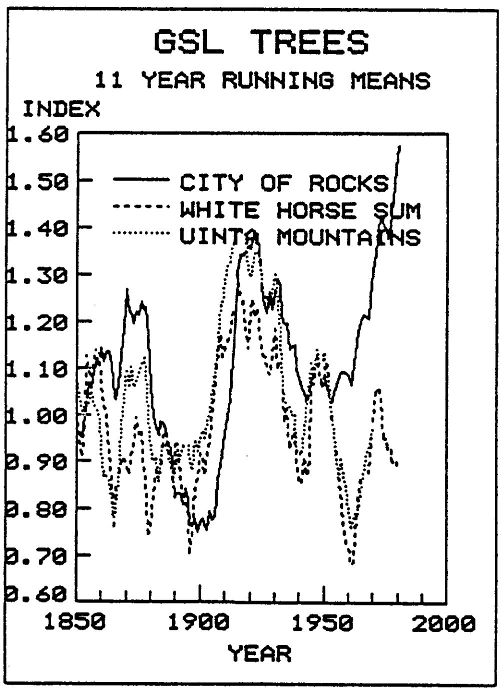


Fig. 2. Plots of 11-year running means of the tree ring index for the three sites near the perimeter of the Great Salt Lake basin that were used in in the studies reported in this paper. An index value of 1.00 represents the average rate of growth for trees at the site.

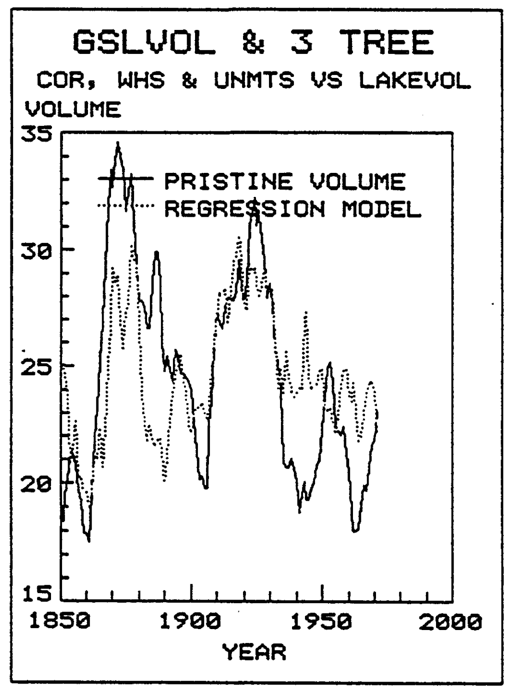


Fig. 3. The relationship between pristine lake volume and a multiple regression model based on all three tree sites over the historic period. The model seriously underestimates the extremes, resulting in an adjusted R^2 of 0.36.

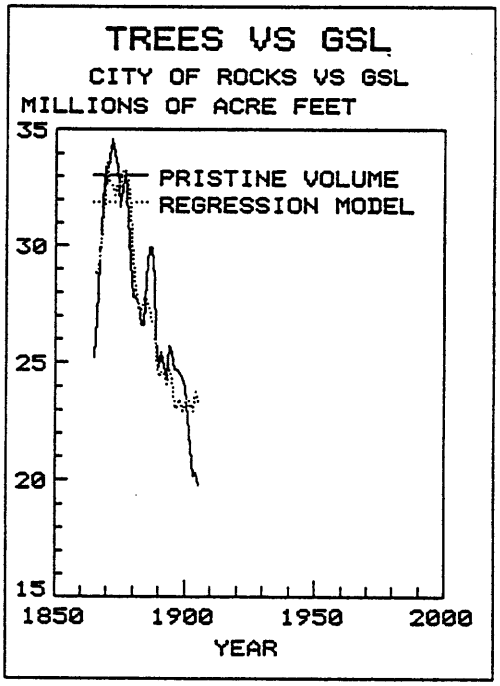


Fig. 4. The relationship between pristine lake volume and tree growth at City of Rocks, Idaho during the period 1865 to 1905. During this period, there is a correlation of 0.83 for the two data sets.

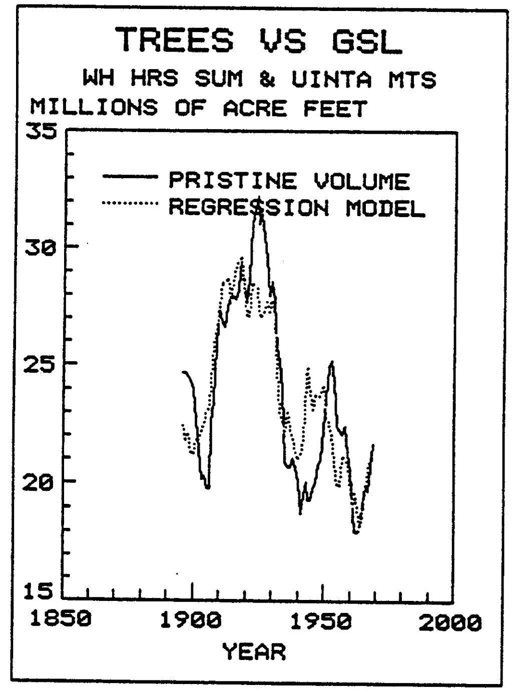


Fig. 5. The relationship between pristine lake volume and a multiple regression model based on tree growth at White Horse Summit, Nevada and in the Unita Mountains of Utah during the period 1905 to 1971. R^2 over the period is 0.68.

GSL basin occurred near the turn of the century. The strong drought that occurred in the northern part of the state at this time was probably associated with this shift. The 1876 lake peak seems to have resulted from heavy precipitation in the northern portion of the GSL basin, probably related to winter & spring frontal storms from the northwest. The 1924 peak appears to have been associated with a generally uniform precipitation pattern, while the 1952 peak is missing from the COR data.

Richardson, 1977 analyzed the combined tree ring data for the Colorado River Basin (CRB) between the period 65 BC and 1970 in an attempt to determine if tree rings could be used to indicate future climate trends. In his analysis, he subjected the tree ring index data to the same procedures used by NOAA to compute climatic normals. Normals are essentially 30-year running means, with a fixed add and drop period of ten years. The resultant "normal" average is plotted at the end point of the averaging period. Based on a plot of "normally" averaged tree ring index data, he concluded that the Colorado River basin was at the end of a severe drought and speculated that the GSL basin could expect considerably greater precipitation in the near future. Since over ten years had lapsed since his original work, we decided to check his analysis using more recently acquired tree ring data from sties which more closely represent the GSL basin. "Normally" averaged tree ring index data were prepared for the three indicator sites between the period 1500 to 1980 (1970 for UMT). These data for the individual sites are shown in Fig. 6, and are grouped and compared to the CRB data in Fig. 7.

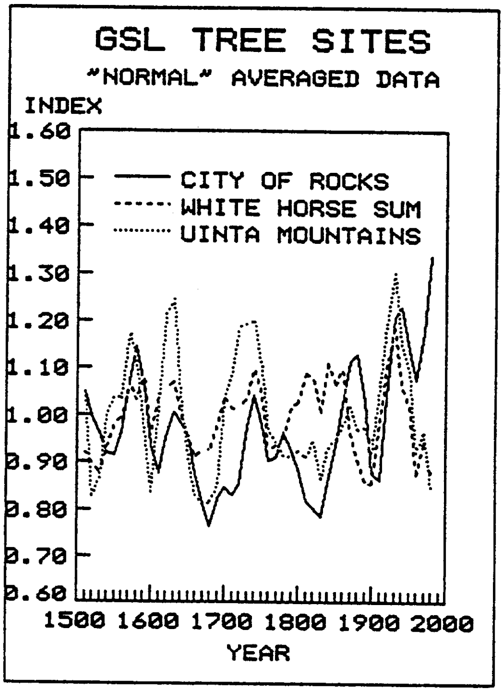


Fig. 6. Data from each tree site was averaged using the same techniques used by the National Weather Service to calculate climatic "normals". These normals are plotted here for the period 1500 to 1980. A repeatable pattern of significant phase shifts between the three sites can be seen. Also of interest is the behavior of the City Of Rocks, ID siteduring the 1900's.

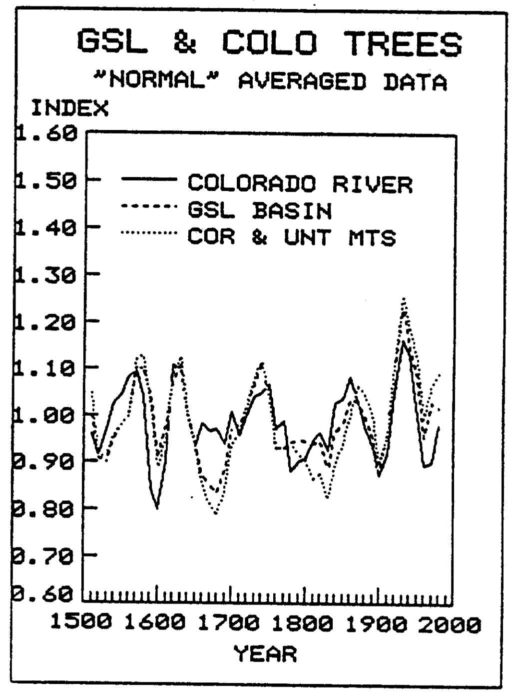


Fig. 7. A comparison of "normal" averaged tree ring data for the GSL basin and for the Colorado River basin. The GSL basin trace is an average of the decade normal values of all three sites shown in Fig. 6. Significant phase differences are also evident in these data. Of additional interest are the reversals in the relative growth rates of the basins during the period.

When data for all three GSL basin sites are averaged (Fig. 7) the climate since 1900 is seen to have been in a wetter pattern than the CRB. At the time of Richardson's original analysis, the normal derived from the three GSL basin sites was very close to their total period average. The higher GSL basin average during the 1900 to 1980 period was largely due to increasingly good growing conditions at the City of Rocks. The GSL basin record does not always show the same relationship to the CRB. During the late 1600's and the mid 1800's, the CRB trees grew much better than the trees to the north and west. The outstanding feature of this data set is the obvious phase lag that exists between the GSL and CRB trees and within the GSL basin alone. Thirty- to fifty-year phase shifts occur in the basin averaged data, and are common between the individual tree sites of the the GSL basin. From this we are led to conclude that wet and dry patterns are shifting over the region with long term persistence.

These persistent wet and dry weather patterns seem to be migratory, and it is possible that their movement could be predictable if subjected to further study. The specific limitation on the use of tree ring data to extrapolate lake levels to previous periods appears to be associated with these moving weather patterns. Since these long term persistent patterns are often smaller than the GSL basin, low predictive power is often the result. An intense wet period in one region can significantly effect the lake as much as a general but more mild pattern. A strong and general pattern, however, appears to always result in a significant lake change. The 1980 - 85 period was of this type. The northern portion of the basin, represented by COR has had an increasing index (with two major cycles) since the early 1800s. When the trees were sampled in 1980, this average stood at an all time high. When this wetter and cooler than normal pattern suddenly shifted sharply to the south in the 1980's, the result was a record increase in lake levels.

CONCLUSIONS:

Based on the previous discussion and analysis, we tender the following conclusions:

1. A somewhat weak, but persistent relationship exists between eleven year running means of the complete historic precipitation records collected near the GSL (Salt Lake City and Corinne data) and lake level or estimated pristine volume. A significantly greater correlation exists for some periods 30 to 50 year subsets of this record. The relationship can be strengthened somewhat by including sites that represent a larger portion of the basin, but the relationship is limited by other known, physical factors. A regression model based on precipitation is strongly strengthened when the previous years lake level is included. In this form the precipitation becomes a strong predictor of the comming years peak lake level.

2. Tree ring index data from sites on the perimeter of the GSL basin, averaged together are an even poorer predictor of lake volume changes than the local precipitation data. However, significant long-term correlations also exist between specific sites and portions of the lake volume trace. Plots of the specific locations with time show that persistent long term (50- to 75-year) patterns of wet and dry cycles rotate around the basin on a scale smaller than the total drainage. When a wet or dry pattern coincides over the whole basin, lake status changes rapidly.

3. Richardson's early analysis of the state of the GSL basin climate, based on Colorado River Basin tree data concluded the climatic pattern then in effect was overly dry, because it did not include the strongly increasing moisture levels of the northern GSL basin. When the persisting wet pattern over southern Idaho moved south to cover the GSL basin, lake levels responded with record jumps.

4. Significant insight into the climatic relationships of the GSL basin and for the rest of the state can be developed by further study of the tree ring and regionalized weather data, including synoptic patterns. The study would be greatly advanced by additional tree ring sampling in the immediate GSL drainage, such as along the Wasatch Front.

5. A recent outlook issued by our office calls for a return to more normal precipitation levels during the next two years, away from the extremely wet conditions observed over the northern portions of the state the last 5 years. However, this outlook should not be interpreted to mean that we have reached the end of the long term trend toward increasing precipitation in the basin. Rather, we believe this is but a return to the more gentile increase observed during the last 20 years. This outlook was based on three factors, a trend analysis of regional precipitation and temperature patterns, the observed return to expected conditions in the Pacific, and the observed northward movement of the wetter than normal precipitation patterns in the southern part of the state.

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Recent Heavy Rains in the Vicinity of the Great Salt Lake:

Just How Unusual?

Thomas R. Karl
Pamela J. Young

National Climatic Data Center
Federal Building
Asheville, NC 28801

Abstract

A long time series (1863-1984) of areal average precipitation in the vicinity of the Great Salt Lake is shown to be highly correlated with the Great Salt Lake levels. This time series is used to assess the unusualness of the recent episode of heavy precipitation (1981 through 1984). The Palmer Drought Severity Index (PDSI) is used to identify wet spells of weather. The cumulative excess precipitation during each wet spell was analyzed using stochastic frequency analysis. The analysis indicates that there were two very important wet spells in the time series, one beginning and ending in the 1860s and the most recent wet spell which began in late 1981. The analysis suggests that the recent heavy precipitation is <u>not</u> unexpected. Assuming the climate remains quasi-stationary over the next 100 years another wet spell can be anticipated to be at least as severe, in terms of excess precipitation, as the 1981-84 wet spell. Whether lake levels can recede to sufficiently low levels to prevent new record high levels during the next severe wet period is uncertain, but it must be considered in risk assessment strategies.

1. Introduction

The modern historical record (1847-present) of the estimated and measured levels of the Great Salt Lake (GSL) indicates a rise in lake levels from record low levels in 1963 to near-record high levels by the end of 1984. Natural lake levels, that is the level the lake would have attained without man's interference, have been estimated at record levels by 1984 (Arnow, 1983, 1985). Sharp increases in lake levels have occurred since 1981. This rapid rise has coincided with several years of very heavy precipitation. The rapid rise in lake level has generated considerable interest because of the practical importance to those living in the vicinity of the GSL (cf. Workshop on the Problems and Prospects for Predicting Great Salt Lake Levels).

Since the lake levels are largely a function of precipitation some interesting questions can be posed pertaining to the recent heavy rains. Some of the questions that arise pertain to whether the recent heavy precipitation is characteristic or uncharacteristic of the recent climate in the GSL area. Three specific questions are addressed in this article. Is the recent heavy precipitation which has led to rapid rises in lake levels characteristic of the climate in the GSL area? If the recent event is a moderately rare event in a quasi-stationary climate, could such heavy rains have been anticipated by a statistical analysis of past data? What inferences can be made regarding future lake levels based on past precipitation records?

2. Data

a) Observed Data

A network of long-term first order and cooperative weather stations was used to assess the recent episode of heavy rains in the GSL area. Fig. 1 depicts the locations of these stations with respect to the GSL. Fig. 2 documents the availability of monthly precipitation data at these stations as well as the timing, number, and distance of station relocations. Camp Douglas is not included in Fig. 2 since monthly precipitation and temperature data were available for only a short period, 1863 through 1882 and from 1889 through 1891. Data from this station (Schott, 1876 and unpublished manuscript records) was used to extend the downtown[1] Salt Lake City time series back in time to 1863.

The estimated and observed peak level of the GSL was used in each of the water years (October through September) from 1863 through 1984. Arnow (1980), describes these data. Direct observations began in 1875 and lake levels were estimated previous to this time. The water year 1984 is defined as October 1983 through September 1984. In addition, the natural lake level was also used at the start of each of these water years. These natural lake levels are estimated by the addition of all known man-made water intake and diversions from the lake (Arnow, 1983).

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[1] We use data from downtown Salt Lake City exclusively. Airport data is not used.

b) Derived Data

The network of stations depicted in Fig. 1 was subdivided into two districts: 1) those stations depicted by dots, referred to as the Salt Lake Drainage, and 2) those stations depicted by triangles, referred to as the Bear River Drainage. An areal mean monthly precipitation total for these two areas was calculated by equally weighting each station's monthly precipitation. If a station was missing or inoperative the normal ratio method (Linsley, et al. 1975) was used to estimate the precipitation that may have occurred if the station was operating. This prevents the network areal average precipitation from becoming biased due to the closing of a station with significantly more or less precipitation than the network mean. For the Salt Lake Drainage the monthly total areal mean average precipitation was calculated from January 1874 through December 1984 and for the Bear River Drainage from January 1903 through December 1984. Additionally, these two networks were combined by equally weighting (weights equal to 0.5) both networks to obtain an areal average precipitation for both drainages for the period January 1903 through December 1984.

Because lake levels were very high during the late 1860s and early 1870s, the areal averages for the Salt Lake Drainage were extended back to 1863. This time series is referred to as Salt Lake Drainage Extended. The procedures used to obtain the areal mean for this network are identical to those just described except that data from Camp Douglas were used to estimate the precipitation that would have fallen at the Salt Lake City station. Simultaneous operation of these two sites, separated by approximately 10 km, occurred during the years 1874 through 1882 and 1889 through 1891. The annual average precipitation at Camp Douglas was within one percent of Salt Lake City's annual average during the years of simultaneous operation. For this reason the precipitation at Camp Douglas during the years 1863 to 1874 was used without adjustment for Salt Lake City. It is noteworthy that for this time series, only one station contributes to the month-to-month and year-to-year variations of precipitation during the years 1863 to 1870.

Fig. 2 indicates a number of significant station moves in excess of 1 km and quite a few station relocations or instrument relocations of shorter distances. For comparative purposes with the calculated areal mean precipitation in the various networks just described, two other methods of computing network average monthly total precipitation were tried using procedures similar to that outlined by Nelson, et al. (1979). These methods are referred to as variants of the modified areal mean index. In the first set, adjustments were made to only those stations with a 1 km or greater documented station relocation. Adjustments to the monthly precipitation were based on the change of the ratio, before and after the relocation, of the annual precipitation at the station with the relocation to the rest of the stations in the network without documented station relocations of 1 km or more. The number of years (y) and stations (n) used to calculate the ratio at the homogeneous stations was determined by:

$$\text{MAX}\ [(\sqrt{n})\ (\sqrt{y})] \quad , \qquad (1)$$

before and after the station relocation at the inhomogeneous station. Sometimes the inhomogeneous station would dictate the use of fewer years than Eq. 1 would suggest because of another station move at the inhomogeneous station before y years had elapsed. In the second set similar adjustments were made whenever there was any documented station relocations regardless of the distance of the station or instrument move.

The Palmer Drought Severity Index (PDSI) (Palmer, 1965) was used to objectively define spells of wet (and dry) weather over the history of our station networks. The PDSI requires monthly mean temperatures for evapotranspiration calculations, but Karl (1985 a,b) indicates that the PDSI is not very sensitive to small changes in the mean monthly temperature. For this reason the temperature used for the calculations in the Salt Lake Drainage was simply that measured at Ogden Pioneer from 1871 to 1984 and at Camp Douglas during the years 1863 to 1870. Based on monthly mean temperatures during simultaneous operation of Camp Douglas and Ogden Pioneer, the Camp Douglas temperatures were adjusted to compensate for the small differences in mean monthly temperatures of the two stations. Laketown's monthly mean temperatures were used in the Bear River Drainage network, and when both the Bear and Salt Lake Drainage networks were combined equal weight was given to Ogden's and Laketown's temperatures. If either Ogden or Laketown had missing monthly temperatures the nearest station was used, after adjusting for differences in the mean, in order to serially complete the time series. Missing values were infrequent.

3. Methodology

The relationship of the various precipitation networks and the various methods of adjusting for station inhomogeneities were correlated with the natural lake levels. Specifically, the departure from the mean annual precipitation was calculated for each water year. A cumulative sum of these departures was derived to produce a time series of cumulative departures from the mean. Cross correlations of the slope of this time series with the slope of the time series of natural lake levels provides a quantitative measure of the relationship between the lake levels and the calculated areal mean precipitation.

After establishing this relationship we address the three questions posed in Section 1. The PDSI was used to identify the beginning and ending times of various wet spells (and dry spells). The sum of the cumulative departures of monthly total precipitation from the mean monthly total precipitation was used as a quantitative measure of the severity of various wet spells. Each wet spell begins when the PDSI exceeds 0.99 (dry spells <-0.99) and ends when it falls below 0.50 (above -0.50 for dry spells). Qualitative descriptions of the magnitude of the PDSI are provided in Table 1. The primary advantages of defining wet spells in this manner are twofold. First, a quantity of precipitation during the recent heavy rainfall regime can be related back to previous heavy rainfall regimes, and second, each wet spell or event is independent of past events. This is particularly important in our quest to answer the questions posed in Section 1 since they require use of stochastic frequency analysis. Kite (1977) thoroughly describes the methodology of stochastic frequency analysis used in this study.

The method of stochastic frequency analysis is entirely data dependent. It can be used to estimate the average return period of various events, i.e., the average length of time between occurrences of an event with a given magnitude. In this regard if an analysis of the recent heavy precipitation event indicates that the recent event is extremely unusual, then we might be tempted to suggest that the recent event may be part of a new climatic regime or that we do not have enough data to adequately characterize such an unusual event. On the other hand, if the event is shown to be only a moderately rare event, then we can be more confident in stating that the recent event is merely an anomaly in the climate record and indeed characteristic of the ongoing climate. Furthermore, we can make future projections regarding the likelihood of similar or even more significant events assuming the climate continues to remain at least quasi-stationary.

In using stochastic frequency analysis there are several assumptions and sources of error involved with an estimation of the average return period of any given event. The important assumptions for stochastic frequency analysis are:

1) The time series is stationary (no climate changes) and
2) Each event is independent of other events.

The important sources of error are:

1) The "true" distribution of the magnitude of each wet spell is unknown
2) There may not currently exist a theoretical frequency distribution to fit the "true" distribution, and
3) Once a probability distribution is chosen, the statistical parameters of the distribution must be estimated.

Each assumption and source of error is discussed in terms of the data from our station network in Section 4b.

4. Results

a) A historical perspective

Fig. 3 depicts the precipitation climate over various networks as calculated using only the normal ratio method to estimate missing data and data from inoperative stations. The precipitation for the Bear River Drainage network depicts two rather well defined peaks of precipitation, the current event beginning in the 1980s and another event in the early 1900s. The most recent period of heavy precipitation is comparable to the event in the early 1900s. There is no evidence to suggest non-stationarity of the time series. When the Bear River and Salt Lake Drainages are combined, a similar scenario arises, but the more recent event is more pronounced than the wet period in the early 1900s. This is attributed to the influence of the Salt Lake Drainage network as Fig. 3c indicates that the recent heavy precipitation of the last few years is substantially greater than any other period since 1875. In fact, since the mid-1960s the filtered time series of annual precipitation is consistently as high or higher than any other time in the record. This is suggestive of non-stationarity in the time series. Further evidence of non-stationarity occurs during the period from 1875 through about 1905, a long period of relatively low precipitation which rather rapidly changes to wetter conditions. Both this long dry spell and the period of wet weather since the mid-1960s suggest that the climate has rather rapidly undergone periods of fluctuation. The extended network for the Salt Lake Drainage (Fig. 3d) depicts a period of heavy precipitation comparable to that observed in recent years. It tends to support the hypothesis that the time series does contain climate fluctuations, but the evidence for a "runaway" change in climate in recent years with new population statistics is not convincing in light of the heavy precipitation in the early 1860s.

The relationship between the precipitation calculated for the four networks and the GSL levels is depicted in Fig. 4. The slopes of the two curves, the cumulative precipitation departures from the mean and the natural GSL levels, appear to have a fairly good match. When the Salt Lake and Bear River Drainage precipitation networks are combined, the relationship to the GSL levels is enhanced, as evidenced by the sign of the slopes of the two curves during the 1930s and 1940s. More important however, the Salt Lake Drainage Extended network (Fig. 4d) depicts the rapid rise in lake levels during the late 1860s at a time when only one station was in the Salt Lake Drainage network. The Salt Lake Drainage Extended network also depicts a long dry spell shortly after the period of very wet weather in the 1860s as well as the period of enhanced precipitation since the 1960s, including the recent three-year period of very heavy precipitation.

The relationship between the running three-year slopes of the cumulative precipitation departures from the mean and the natural levels of the lake are quantified in Table 2. Three years is depicted since this roughly corresponds to the length of the recent heavy precipitation episode. The correlations are largest at a lag of one reflecting the delay in the response of the lake. All correlations are relatively high at the first and second lags, but the Bear River and Salt Lake Drainages combined have the highest correlations. The Salt Lake and Bear River Drainages have similar correlations to the GSL levels for identical periods of record, but not when the Salt Lake Drainage is extended further back in time. This undoubtedly is attributable to the fewer stations in the network during the nineteenth century and the lack of measured lake levels, i.e., estimates are used. Generally, the results suggest that the precipitation networks used in this study relate well to the fluctuations of the lake levels. When the areal average precipitation was calculated by each of the two variants of the modified areal mean index the correlations were quite similar to those reported in Table 2. This suggests that the station relocations in our network do not introduce significant bias into the time series.

Fig. 5 depicts the PDSI for each of the four networks. The recent wet spell is highlighted in the Bear River Drainage by large positive PDSIs, but values nearly as large are also depicted in the early 1900s. When both the Bear River and the Salt Lake networks are combined, the recent event is amplified, similar to that depicted in Fig. 3b. Little evidence of non-stationarity is apparent in Fig. 5b. Figs. 5c and 5d however, depict a long period of severe and extreme drought during the late 1800s. This immediately follows the very high values of PDSI in the 1860s depicted in Fig. 5d. The Salt Lake Drainage Extended time series indicates that the climate may be stationary about some long-term mean, but for shorter periods, fluctuations about this mean can and do occur.

In Fig. 6a the impact of the severe droughts and wet spells (PDSI $\geq |3.0|$) on the cumulative excess or deficiency of water during each spell is clearly evident. The dry weather of the late 1800s is seen as largely the result of two extended dry periods which resulted in a prolonged decrease in lake levels. The 1860s wet spell in the Salt Lake Drainage Extended network is comparable, in terms of cumulative excess precipitation, to the current wet spell. Again, a rather strong argument can be made for the existence of climate fluctuations in the Salt Lake Drainage area as evidenced by the prolonged droughts in the 1800s. It is also noteworthy that the observed lake levels in Fig. 6 rather closely correspond to the natural levels depicted in Fig. 4.

b) Stochastic frequency analysis

We limited our frequency analysis to the Salt Lake Drainage Extended network because it has the longest record, and yet, the cumulative departure of the precipitation from the mean closely resembles natural and observed lake levels. Table 3 lists all the wet periods as defined by the PDSI, and their associated cumulative excess precipitation. By definition, each of these events are independent, and as a result our requirement for independence of events is satisfied.

We tested for stationarity of the events using the relatively powerful non-parametric Mann and Whitney U test. The 48 events listed in Table 3 were divided into two groups of equal years (not events). The first group contained events of cumulative precipitation from 1863 through 1923 and the second group the remaining years. The first set contained only 15 wet periods compared to 33 wet periods in the second subsample. For this reason when the Mann and Whitney U test was performed, the two samples were significantly different at the 99% significance level. As we have indicated, this is primarily attributable to the long drought in the late 1800s. In order to find a sample that meets the stationarity assumption we are forced to limit our analysis to the 21-year period 1964 to 1984. This can be done, and we will include return period estimates based on this small sample. But in addition, we will fit theoretical distributions to much larger samples, i.e., 1863 thru 1984. We argue that despite the periods of apparent non-stationarity it is quite possible that they are manifestations of climate fluctuations which irregularly fluctuate about a much longer stable mean, i.e., quasi-stationary. Evidence for such a contention is the abrupt change from an extremely wet spell in the 1860s to a prolonged drought in the late nineteenth century.

Using the cumulative excess precipitation departures from the mean the truncated normal, the lognormal, the 3-parameter lognormal, the Gumbel, the Pearson Type III, and the log-Pearson Type III were all fit to the event magnitudes for six sample sizes, 1863 through 1984, 1863 through 1976, 1863 through 1976, 1875 through 1976, 1964 through 1981, and 1964 through 1984. The lognormal distribution is the most appropriate distribution based on least squares estimates of the observed and predicted event magnitudes and the Chi-Square goodness of fit test. Each distribution was fit using maximum likelihood procedures (except for log-Pearson Type III where the method of moments was required). The maximum likelihood parameter estimation is the most efficient parameter estimator compared to the method of moments, graphical methods, or least squares (Kite, 1977).

The return periods of the various event magnitudes for cumulative excess precipitation in Table 3 are given in Table 4 for various periods of record. The return period estimates of particular interest to us are those associated with the 1,027mm cumulative precipitation excess.

Since our data is neither an annual series (i.e., the highest event in any given year) nor is it a partial series (i.e., multiple events per year, but the number of events used in the frequency analysis is equal to the number of years of data) the standard return period calculations are not applicable. We have modified the procedure used to calculate return periods for an annual series by determining the ratio (R) of the number of years of record to the number of events. This value of R is then used to calculate the appropriate standard normal probabilities (γ) and their standard normal deviates associated with each of the various return periods (T):

$$\gamma = (R/T) - 0.5 \qquad (2)$$

Using the entire period of record, the recent heavy precipitation event has a return period of about 118 years compared to a 146-year return period using data from 1863 through 1976 and 290 years using data from 1875 through 1976 (Table 4). By omitting the added information from Camp Douglas, Corinne, Ogden, and Salt Lake City during the years 1863 through 1874 the return period of the recent event is calculated to be 150 years (third row Table 4). This return period is nearly equal to that obtained by using the full period of record. The consistency of the return period estimates using data from the three periods of record 1863-1984, 1863-1976, and 1875-1984 suggests that the frequency analysis is not highly sensitive to the omission or inclusion of either one of the two heaviest precipitation episodes. All of these return periods are derived from a quasi-stationary time series. When the stationary series from 1964 through 1984 is used the return period estimate for the recent event is 188 years, but when data for the years 1964 through October 1981 is used the return period of the recent event is much greater, 25,000 years. At first, it may seem surprising that the stationary time series with their more frequent wet spells and higher logarithmic means of the event magnitudes usually have longer return periods than the quasi-stationary time series with their lower frequency of wet spells and lower logarithmic means. This is attributable to the variance of logarithms of the event magnitudes. Despite the lower means, the variance is substantially larger in the quasi-stationary time series, particularly whenever either one of the two severe wet spells (1860s and 1980s) is included in the parameter estimation for the lognormal frequency distribution.

If the climate of the GSL can be considered as consisting of a series of short-term climate fluctuations, all fluctuating about a stable mean, then the return period estimates using the entire period of record are preferred. If the climate has truly changed to a new semipermanent state with new population statistics then even the return periods of the stationary time series may be incorrect because of the shortness of the climate regime (1964-84). Based on the behavior of the past record, the assumption of a quasi-stationary climate records seems most appropriate.

If we had calculated our return periods in 1976 using data from 1863 we would have predicted a return period for the most recent event at 146 years. This would certainly qualify as an unusual event, but <u>not</u> uncharacteristic of the climate in the GSL area. Using some basic concepts from probability, risk analysis, and the values of T, we can obtain the probability of not having an event of a given magnitude during any interval of time. Using data from 1863 through 1976 Table 5 indicates that we would have expected a 71% probability of <u>not</u> having a wet spell as extreme as the current wet spell in any 100-year period. The risk of <u>having</u> such an extreme event in any 100-year period however, 29%, is a fairly high risk. In fact using the full period of record (1863-1984) the return period for the recent heavy precipitation event suggests that there is greater than a 50% chance of having a 100-year period with a wet spell as severe as the current spell. Table 5 demonstrates the value of a long time series in projecting the probability of unusual events. The inclusion of the early severe wet spell as well as the less severe wet spells up through the 1970s provides us with much better information regarding the unusualness of the most recent wet spell compared with the shorter period of record 1964-81. In fact, using the data either from any of the three periods 1863 through 1976, 1875 through 1976, or 1863 through 1984 we can state rather unequivocally that the 1981-84 wet spell was overdue.

Obviously, at this date (March 1985) it is uncertain whether the current wet spell is over. Precipitation in the GSL vicinity during both January and February of 1985 was below normal, and the PDSI has been dropping since its peak value in December 1983. Whether this wet spell has ended or will last another year or two, it is unlikely that the recommended return period estimates would change substantially as indicated by a comparison of the event magnitudes of various return periods using data from 1863 through 1976 (i.e., omitting the recent event) and 1863 through 1984 (i.e., including the recent event).

5. Conclusion

The recent heavy precipitation event in the vicinity of the GSL is estimated to recur about once in every 120 years. As such, an event of the magnitude of the recent wet spell <u>is</u> characteristic of the past climate. Assuming the climate remains quasi-stationary it is more likely than not that another event of similar magnitude will occur within the any future 100-year period, but it is not likely (probability < 36%) in any 50-year period. In the intervening time, it is uncertain whether a prolonged dry spell will again strike as in the late 1800s to bring the lake levels down to much lower levels before the next severe wet spell occurs. In terms of practical importance to the local community in the

GSL area the possibility of another severe wet spell before the lake can naturally recede should not be dismissed.

Acknowledgments

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Table 1. - Classes for wet and dry periods

PDSI	Class
≥ 4.00	Extremely wet
3.00 to 3.99	Very wet
2.00 to 2.99	Moderately wet
1.00 to 1.99	Slightly wet
.50 to .99	Incipient wet spell
.49 to -.49	Near normal
-.50 to -.99	Incipient drought
-1.00 to -1.99	Mild drought
-2.00 to -2.99	Moderate drought
-3.00 to -3.99	Severe drought
≤ -4.00	Extreme drought

Table 2. Correlations of the 3-year slope of the natural lake levels on Oct. 1 of each year and the 3-year slope of the cumulative departures from the mean annual precipitation during the water year ending on October 1. Numbers in parenthesis indicate number of data points used in calculating the correlations.

Lag	Salt Lake Drainage Extended	Salt Lake Drainage	Bear River Drainage	Salt Lake Drainage	Bear River and Salt Lake Drainage
0	0.76 (121)	0.76 (109)	0.78 (81)	0.77 (81)	0.83 (81)
1	0.76 (120)	0.78 (108)	0.82 (80)	0.81 (80)	0.86 (80)
2	0.58 (119)	0.60 (107)	0.60 (70)	0.62 (79)	0.65 (79)
3	0.37 (118)	0.38 (106)	0.33 (78)	0.41 (78)	0.39 (78)
4	0.25 (117)	0.25 (105)	0.10 (77)	0.28 (77)	0.20 (77)
5	0.20 (116)	0.18 (104)	-0.04 (76)	0.23 (76)	0.08 (76)

Table 3. Length of wet spells for the Salt Lake Drainage Extended network as defined by the PDSI, and their associated excess precipitation during the length of the wet spell.

Begin Year of Event	End Year of Event	Highest PDSI and (rank)	Excess Precipitation (mm) and (rank)
1864	1865	4.0 (11)	168 (11)
1865	1868	7.9 (2)	869 (2)
1869	1870	2.5 (22)	157 (14)
1875	1876	3.3 (16)	131 (19)
1889	1890	4.6 (5)	195 (8)
1904	1904	1.8 (32)	67 (26)
1906	1907	4.8 (4)	273 (4)
1908	1910	4.9 (3)	361 (3)
1912	1912	1.6 (34)	41 (38)
1913	1914	2.0 (28)	32 (41)
1916	1917	2.9 (19)	90 (22)
1919	1919	1.4 (42)	12 (45)
1920	1920	1.1 (47)	15 (44)
1920	1921	1.8 (31)	55 (31)
1923	1923	1.5 (36)	50 (32)
1924	1925	2.9 (20)	115 (20)
1927	1927	1.2 (45)	45 (35)
1929	1929	1.2 (44)	28 (42)
1929	1929	1.1 (48)	46 (34)
1930	1930	2.7 (21)	114 (21)
1932	1932	2.2 (25)	45 (36)
1935	1935	1.5 (37)	39 (39)
1936	1936	1.7 (33)	56 (30)
1937	1937	1.8 (30)	32 (40)
1938	1938	1.3 (43)	65 (27)
1938	1939	1.4 (41)	2 (47)
1940	1942	3.8 (13)	207 (6)
1944	1944	2.1 (26)	87 (23)
1945	1945	3.2 (17)	135 (18)
1946	1947	3.0 (18)	170 (10)
1949	1949	1.6 (35)	48 (33)
1949	1950	1.5 (38)	1 (48)
1951	1952	4.5 (6)	142 (15)
1953	1953	1.2 (46)	3 (46)
1956	1956	1.9 (29)	23 (43)
1957	1957	2.3 (23)	61 (29)
1961	1962	2.1 (27)	65 (28)
1964	1965	3.7 (14)	141 (16)
1967	1967	4.0 (10)	83 (24)
1968	1969	3.8 (12)	167 (12)
1970	1972	4.5 (7)	247 (5)
1972	1974	3.6 (15)	187 (9)
1975	1976	4.5 (8)	205 (7)
1977	1977	1.4 (40)	42 (37)
1978	1979	2.3 (24)	140 (17)
1980	1980	4.1 (9)	166 (13)
1981	1981	1.5 (39)	74 (25)
1981	1984*	11.0 (1)	1027 (1)

*Wet spell ongoing as of December 1984.

Table 4. Return periods for various values of cumulative excess precipitation (mm) departures from the mean (P) during wet spells and their standard error (s_P). Values are underlined which reflect the cumulative excess precipitation of the most recent wet spell (October 1981 through December 1984).

Salt Lake Drainage Extended

Period of Record		Return Period (Years) 50	100	118	146	150	188	200	250	290	300	25,000
1863-1984	P	610	931	<u>1027</u>	1150	1168	1316	1356	1517	1632	1656	
	s_P	98	153	<u>169</u>	190	194	219	226	253	273	277	
1863-1976	P	548	832	915	<u>1027</u>	1043	1175	1211	1358	1460	1489	
	s_P	77	115	127	<u>142</u>	144	161	166	186	199	203	
1875-1984	P	553	829	910	1006	<u>1027</u>	1153	1189	1330	1426	1456	
	s_P	89	136	150	167	<u>171</u>	192	198	222	239	244	
1875-1976	P	424	618	673	749	758	842	870	963	<u>1027</u>	1042	
	s_P	46	65	70	77	78	85	88	96	<u>102</u>	103	
1964-1984	P	658	840	887	949	957	<u>1027</u>	1048	1119	1169	1179	
	s_P	355	523	569	633	644	<u>716</u>	739	818	875	887	
1964-1981	P	346	407	422	441	444	465	471	492	506	509	<u>1027</u>
	s_P	60	76	79	84	85	91	92	98	102	102	<u>247</u>

Table 5. Probability of non-occurrence of wet spells with cumulative excess precipitation departures from the mean greater than or equal to 1,027mm for the Salt Lake Drainage Extended network.

PERIOD OF RECORD FOR FREQUENCY ANALYSIS	50 YEARS	100 YEARS
1863-1984	0.653	0.427
1863-1976	0.709	0.503
1875-1984	0.716	0.512
1875-1976	0.841	0.708
1964-1981	0.998	0.996
1964-1984	0.765	0.587

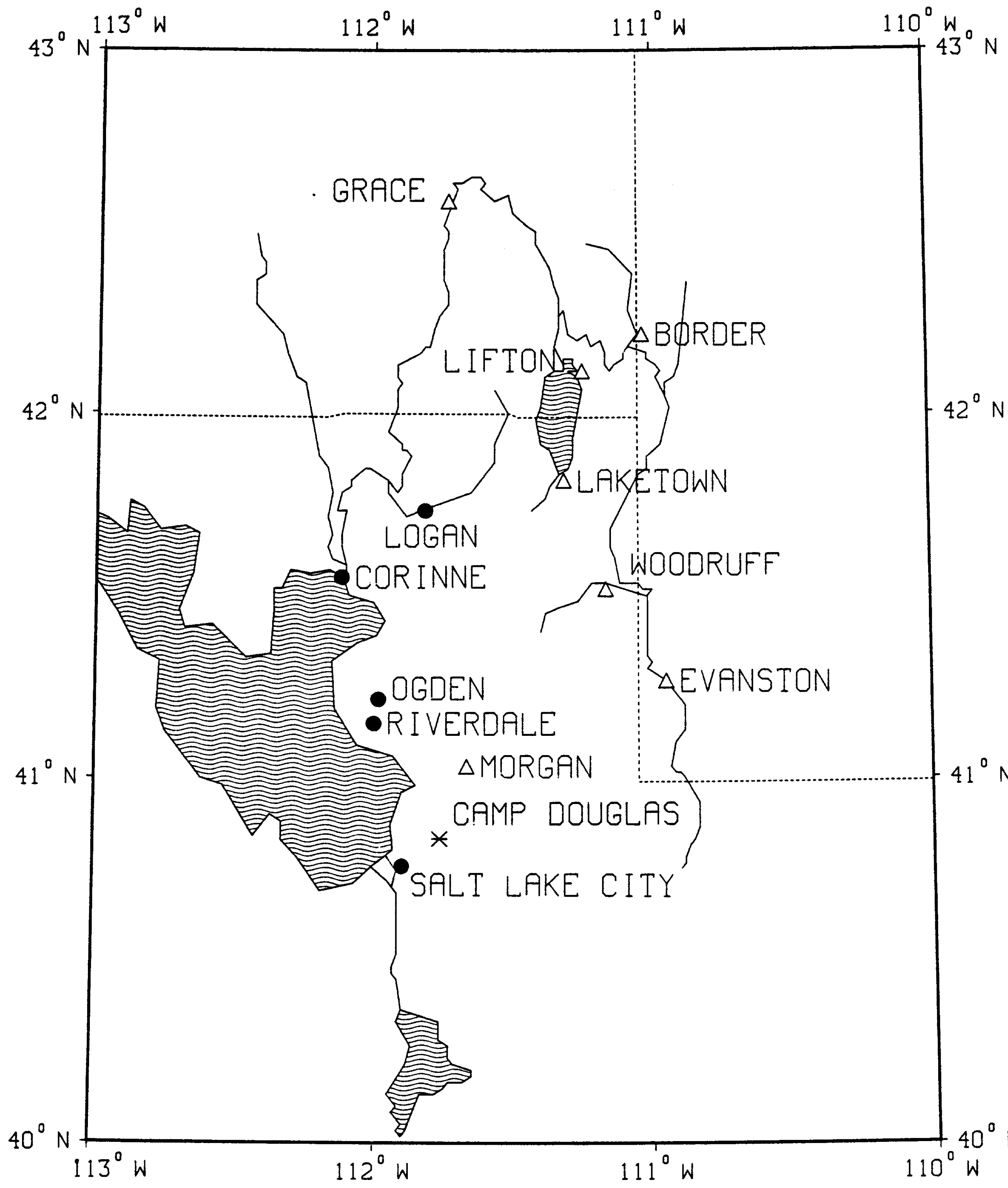

Fig. 1 Station locations for various precipitation networks. Circles denote the Salt Lake Drainage network, triangles the Bear River Drainage, and the asterisk is part of the Salt Lake Drainage Extended network.

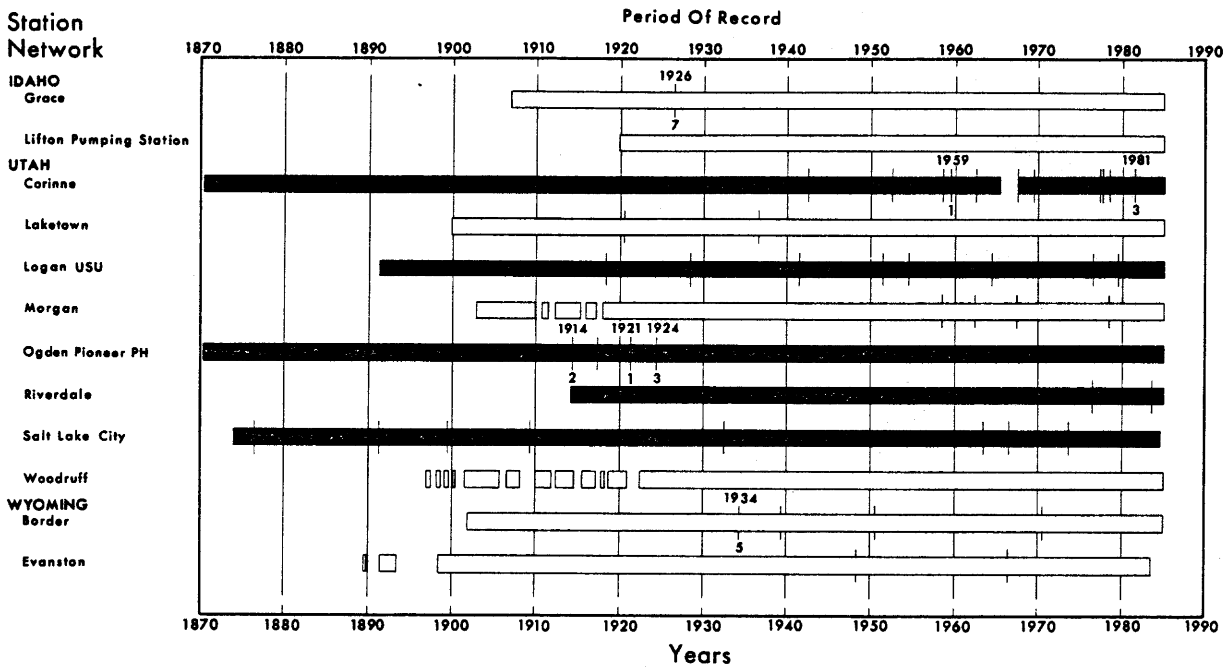

Fig. 2 Station histories for stations used in the various networks. Each tick without a number represents a station relocation of less than 1 km. Numbers beneath ticks represent the distance of station relocations in km. Solid bars denote stations in the Salt Lake Drainage and open bars denote the Bear River Drainage.

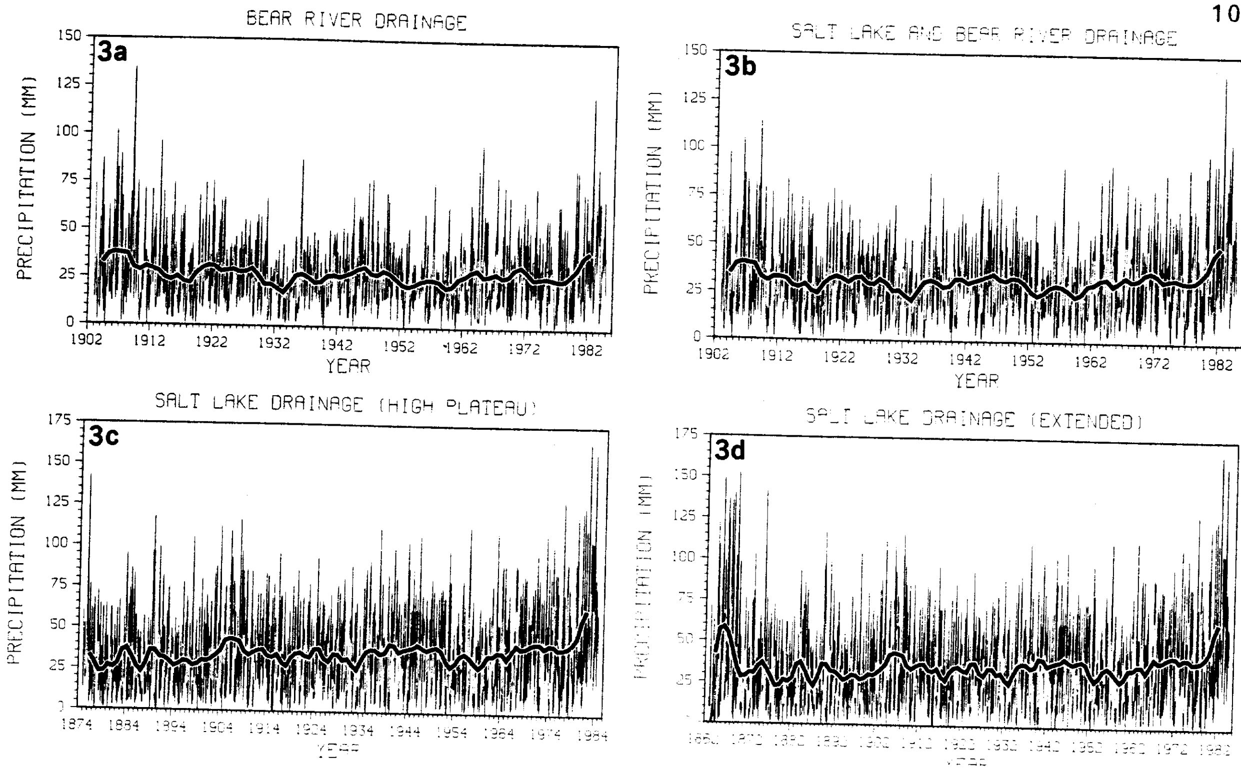

Fig. 3 Monthly areal average total precipitation. Thick curve is a three-term binomial filter (weights of 0.25, 0.50, and 0.25) of the annual monthly average total precipitation.

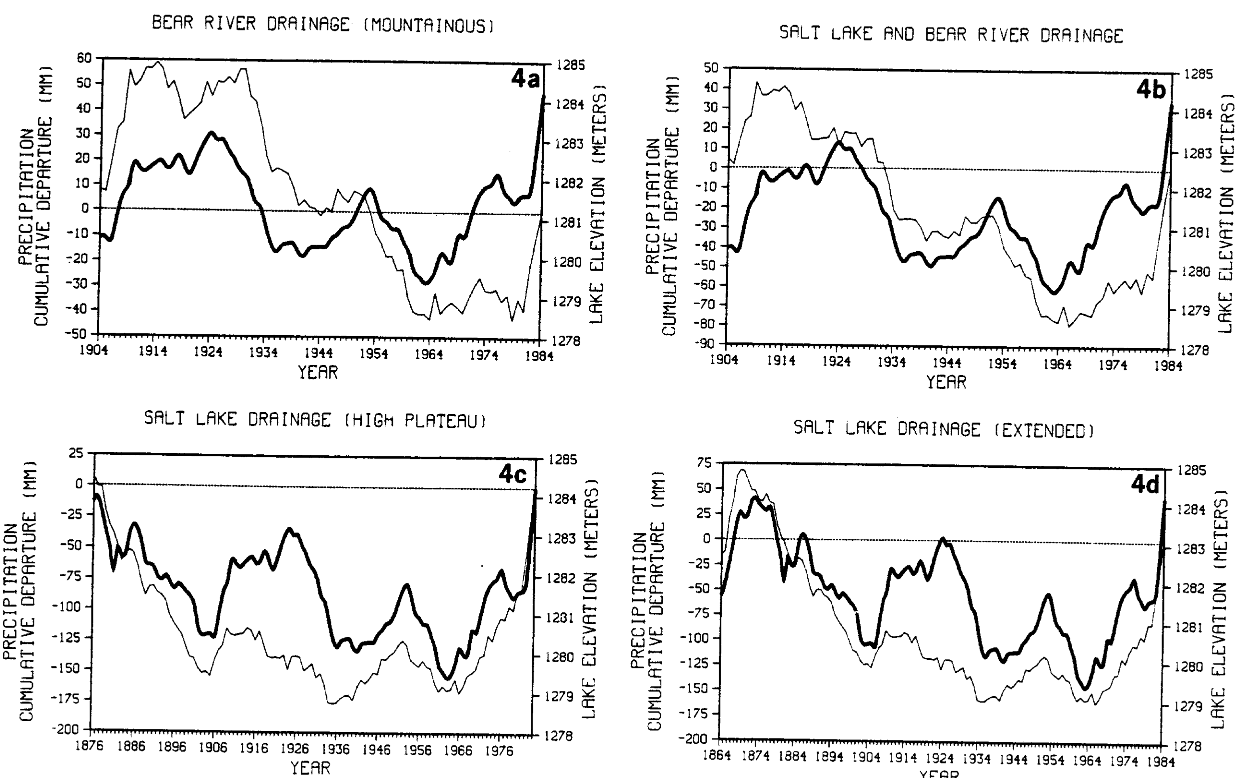

Fig. 4 Salt Lake levels (natural) on Oct. 1 (thick curve) and cumulative departures from the mean water-year precipitation (thin curve).

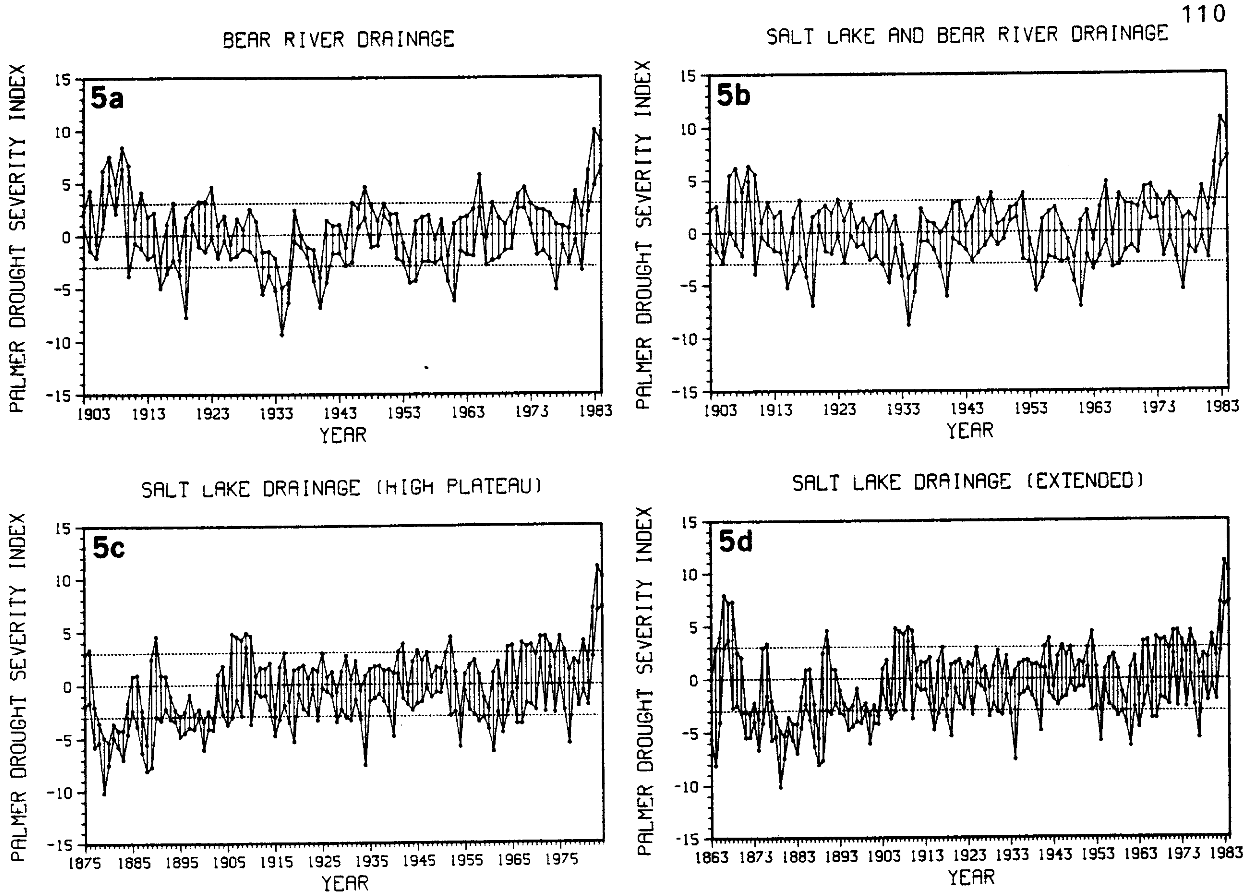

Fig. 5 Annual maximum, minimum, and range of the PDSI. Dashed lines denote near normal conditions (value of 0) and severe droughts or wet spells (values of ± 3).

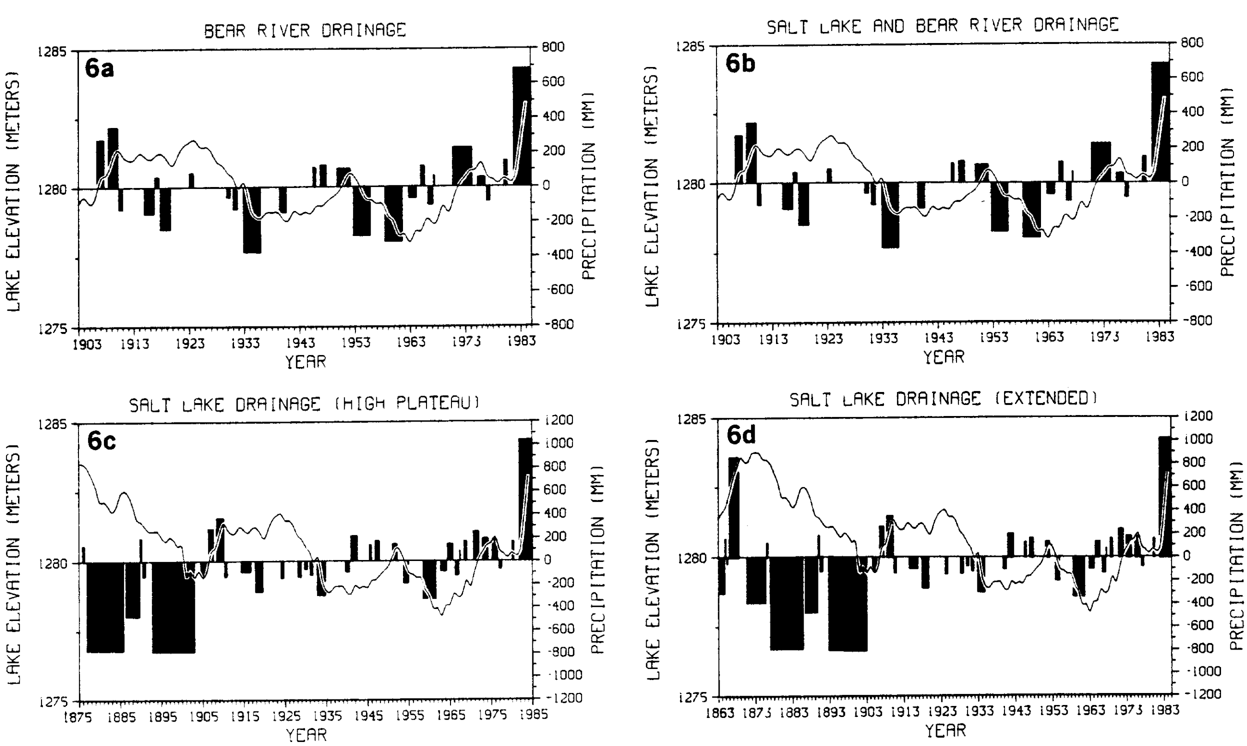

Fig. 6 Observed peak levels of the Great Salt Lake for the water-year ending on September 30 (thin curve) and the cumulative departures from the long-term mean precipitation (bar graphs) during severe wet and dry spells (PDSI $\geq$3.0)

SECULAR CLIMATIC FLUCTUATIONS IN THE GREAT SALT LAKE BASIN

by
J.K. Eischeid, R.S. Bradley and Shao, Xuemei
Department of Geology and Geography
University of Massachusetts, Amherst, Ma.

Introduction

The extraordinarily rapid rise of the Great Salt Lake over the past few years has focused attention on long-term climatic variability in the region. How unusual has the recent period been? Were there any comparable episodes in the past? What are the long-term trends of precipitation and temperature in the area? To answer these and other questions concerning long-term climatic fluctuations, a set of high quality long-term temperature and precipitation station records (Bradley, 1982) was selected from within the drainage basin of the Great Salt Lake (Figure 1, Table 1). More than half of these records exceed 70 years in duration, thereby providing a long-term perspective on contemporary climatic conditions in the area.

In order to compare records from stations which vary widely in elevation, monthly values were expressed as percentages of the 1921-70 mean (for precipitation) or departures from the same period mean (for temperature). The resulting series of anomalies were then averaged, monthby month, to provide a regional index of temperature and precipitation for the region over the last 90 years. Finally, monthly anomalies were used to compute seasonal anomalies using conventional seasonal definitions (i.e. Winter = December, January, February; Spring = March, April, May; Summer = June, July, August; Fall = September, October, November).

Results

Long-term changes of seasonal precipitation (as a percentage of the 1921-70 reference) are shown in Figure 2*. The anomalous nature of the last few years is clearly apparent, particularly in summer and fall months; in 1982, summer and fall precipitation was more than 200% of the long-term mean and in 1983, fall precipitation averaged 323% of "normal". In addition, precipitation in spring months (the wettest season) has been considerably above average since 1980. However, winter precipitation has not been exceptional; in the last decade, only the winter of 1979-80 was unusually wet. Overall, seasonal precipitation anomalies in the last few years have been quite extraordinary in comparison with the last 90 years (Table 2). This has resulted in very high annual or water year (October through September) precipitation totals(figure 3). It is clear, however, that precipitation has been generally increasing since the early 1930s, punctuated by a drier

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*The mean contribution (1921-70) of each season to the annual precpitation total for all stations is as follows:
Spring: 30% Summer: 19% Fall: 23% Winter: 25%

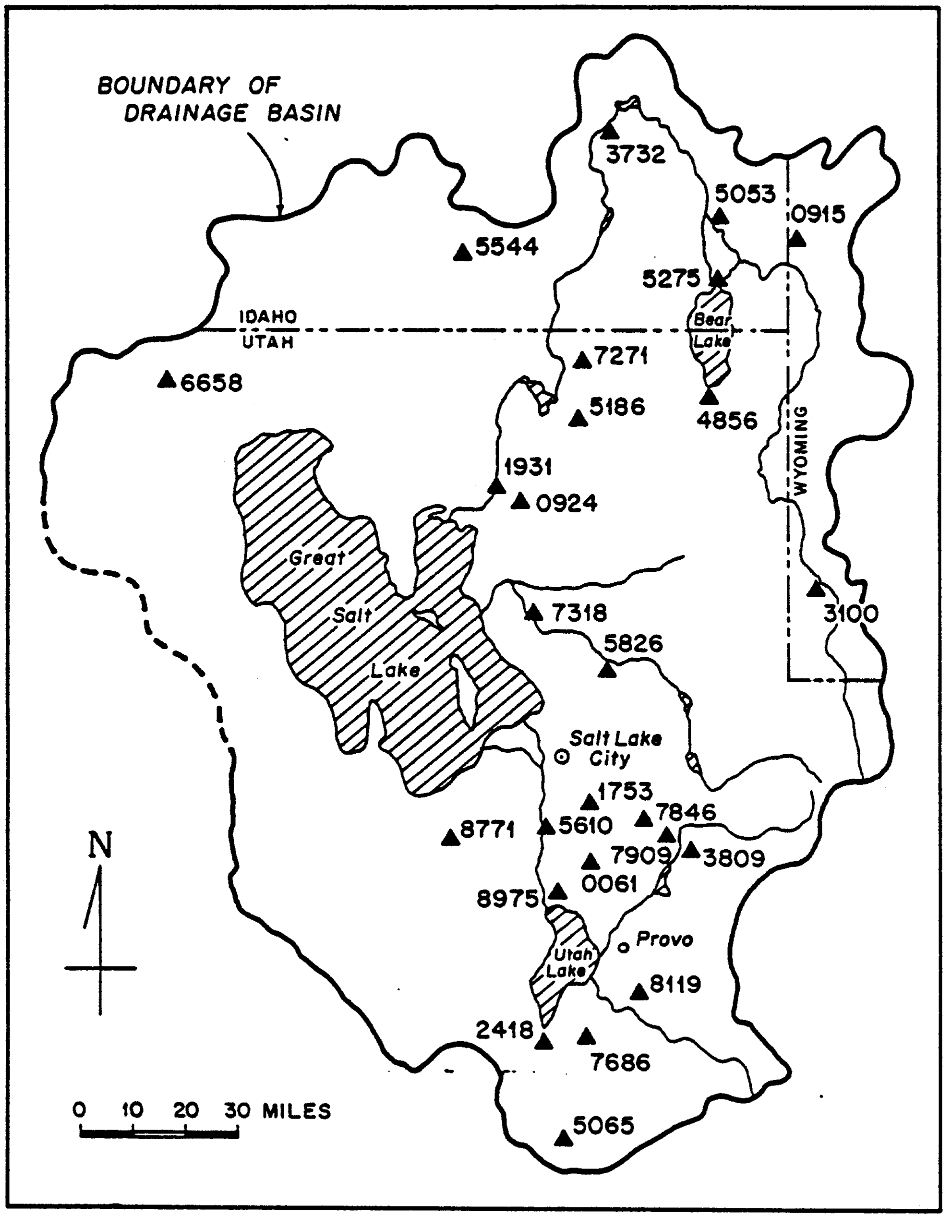

Figure 1. Location of long-term climatic stations used in study in relation to Great Salt Lake Drainage Basin (see Table 1).

Table 1
Climatic Stations Used in Composite Indices

A. Precipitation Stations

Idaho

3732	Grace	1907-Oct. 1984
5275	Lifton	1919-Oct. 1984
5544	Malad	1917-Oct. 1984
6053	Montpelier	1914-Oct. 1984

Utah

0061	Alpine	1910-Oct. 1984
1759	Cottonwood Weir	1917-Oct. 1984
2418	Elberta	1902-Oct. 1984
3809	Heber	1893-Oct. 1984
4856	Laketown	1900-Oct. 1984
5065	Levan	1889-Oct. 1984
5186	Logan (Utah State)	1891-Oct. 1984
5610	Midrale	1911-1978
5826	Morgan	1903-Oct. 1984
6658	Park Valley	1911-Oct. 1984
7271	Richmond	1911-Oct. 1984
7318	Riverdale	1914-Oct. 1984
7686	Santaquin	1914-Oct. 1984
7846	Silver Lake Brighton	1915-Oct. 1984
7909	Snake Creek	1914-Oct. 1984
8119	Spanish Fork P.H.	1910-Oct. 1984
8771	Tooele	1896-Oct. 1984
8973	Utah Lake Lehi	1904-Oct. 1984

Wyoming

0915	Border	1902-Oct. 1984
3100	Evanston	1899-Oct. 1984

B. Temperature Stations

Idaho

3732	Grace	1907-Oct. 1984

Utah

0924	Brigham City	1915-1970
1731	Corrine	1897-Oct. 1984
2418	Elberta	1914-Oct. 1984
3809	Heber	1893-Oct. 1984
4856	Laketown	1912-Oct. 1984
5065	Levan	1891-Oct. 1984
5186	Logan (Utah State)	1891-Oct. 1984
5826	Morgan	1905-Oct. 1984
7318	Riverdale	1914-Oct. 1984
7686	Santaquin	1914-Oct. 1984
7909	Snake Creek	1914-Oct. 1984
8119	Spanish Fork P.H.	1911-Oct. 1984
8771	Tooele	1896-Oct. 1984
8973	Utah Lake Lehi	1914-Oct. 1984

Wyoming

0915	Border	1902-Oct. 1984
3100	Evanston	1898-Oct. 1984

GREAT SALT LAKE PRECIPITATION DEPARTURES FROM 1921-70 MEANS

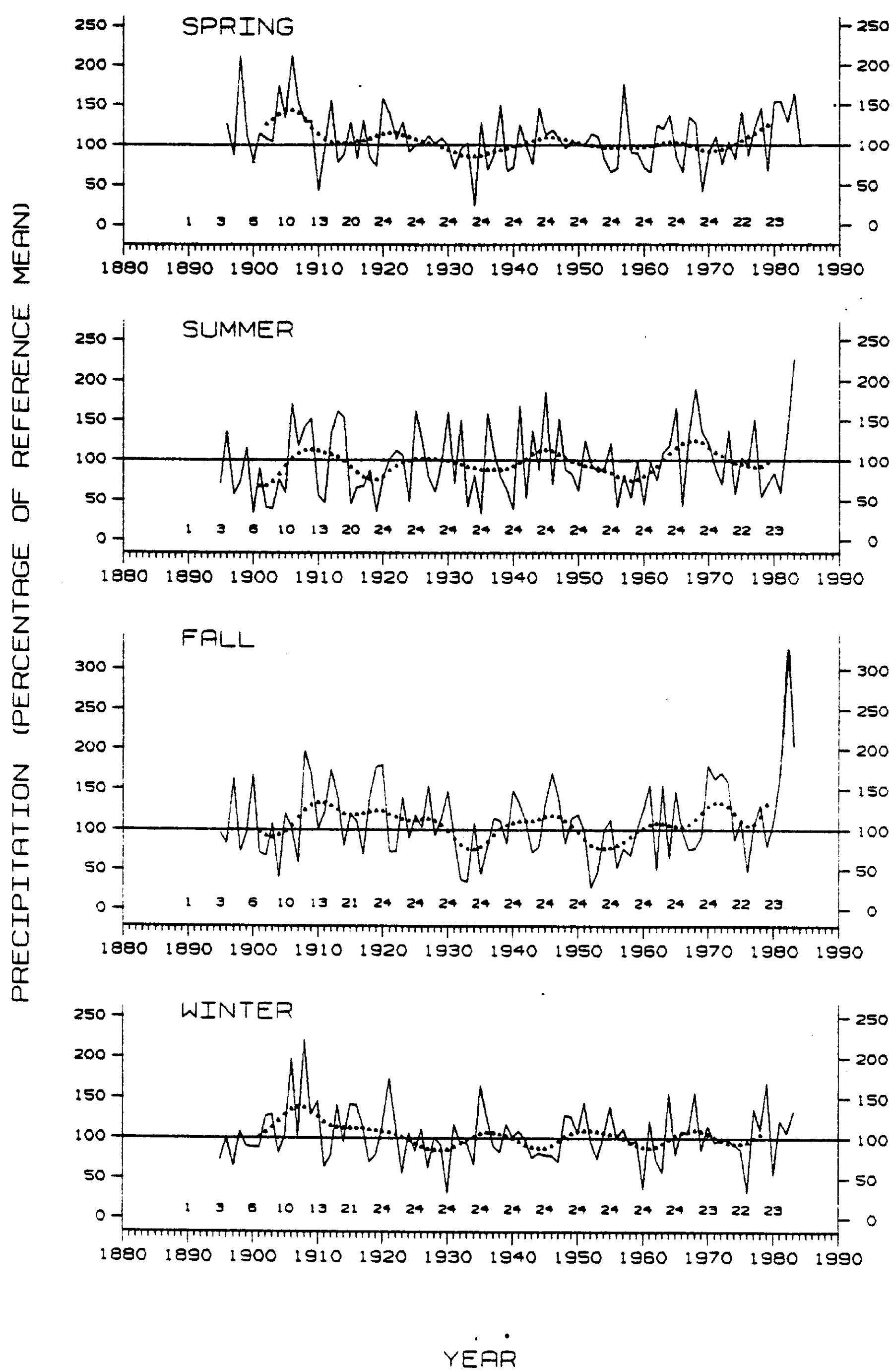

Figure 2. Seasonal precipitation amounts expressed as a percentage of the 1921-70 period. Anomalies calculated for each station in Table la and averaged for the region. Number above abcissa indicates number of station records used in composite average; no value was plotted unless at least 3 station records were available. Data end in 1983 (for spring, summer and fall) and 1983-4 for winter (spring = M,A,M; summer = J,J,A; fall = S,O,N; winter = D,J,F; winter 1983-4 plotted as 1983, etc.). Low frequency trends (shown by darker line) derived from an 11-point binomial filter.

Table 2

10 Wettest

Spring

Rank	Year	Mean(%)
1	1906	212.80
2	1898	212.17
3	1895	182.23
4	1957	176.73
5	1904	175.33
6	1983	164.17
7	1920	157.73
8	1912	156.23
9	1981	154.63
10	1980	154.10

Summer

Rank	Year	Mean(%)
1	1983	226.47
2	1968	190.50
3	1945	185.33
4	1906	171.67
5	1941	169.07
6	1965	165.57
7	1925	162.97
8	1913	161.23
9	1930	160.60
10	1936	159.50

Autumn

Rank	Year	Mean(%)
1	1982	323.27
2	1983	204.50
3	1908	197.60
4	1920	180.50
5	1970	180.07
6	1919	177.90
7	1912	174.73
8	1972	170.70
9	1946	170.33
10	1909	170.03

Winter

Rank	Year	Mean(%)
1	1908-09	220.87
2	1906-07	197.87
3	1921-22	174.07
4	1979-80	169.43
5	1935-36	165.23
6	1968-69	156.27
7	1964-65	155.77
8	1910-11	145.53
9	1951-52	144.17
10	1915-16	142.27

Annual

Rank	Year	Mean(%)
1	1983	190.52
2	1909	176.19
3	1982	172.98
4	1906	153.52
5	1941	135.07
6	1907	134.63
7	1908	134.45
8	1980	132.03
9	1912	131.83
10	1945	131.70

Water Year(October-September)

Rank	Year	Mean(%)
1	1981-82	188.2
2	1908-09	172.2
3	1982-83	166.8
4	1983-84	164.7
5	1905-06	142.8
6	1972-73	137.7
7	1946-47	136.6
8	1964-65	135.0
9	1906-07	134.9
10	1979-80	131.0

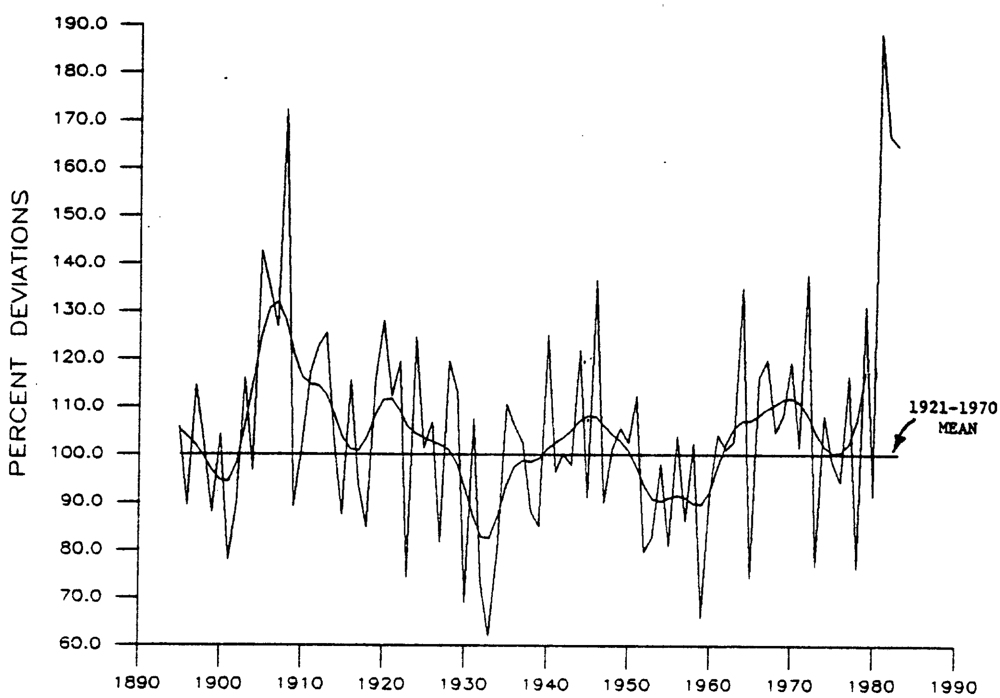

Figure 3. Water year (October through September) precipitation anomalies, 1895-96 to 1983-84 averaged over the Great Salt Lake drainage basin.

episode in the 1950s and to a lesser extent in the mid-1970s. Precipitation amounts earlier this century appear to have been generally higher (though fewer data are available prior to 1915). Nevertheless, the only sequence of years which appear to have been comparable with the recent period was 1905-1908.

Analysis of long-term mean temperature records for the Great Salt Lake basin does not reveal anything particularly unusual about the last few years (Figure 4). In particular, summer temperature anomalies (which influence evaporation rates over the lake) have been close to the long-term mean for more than a decade. Temperatures generally increased to a peak centered around 1950 and have been declining in all seasons, except winter, since then (Figure 5a). This is the opposite of the general low frequency trend observed in the precipitation data (Figure 5b); in fact, there is a statistically significant negative correlation between seasonal temperature and precipitation anomalies (except in winter months)(Table 3).

Spectral analysis of the records shows a few statistically significant periodicities, though none have sufficient power to be of predictive value. The most statistically significant periodicities are shown in Table 4; detrending of the data did not alter the results. Of particular interest is the 2 to 2.2. periodicity in summer precipitation. Previous work in the southwestern United States (Sellers,1960) and in the Rocky Mountain States (Bradley, 1976) indicates that this periodicity is of regional significance (Figure 6). It has also been noted in soil moisture recharge, snowpack and streamflow data by Engelen (1972). Reasons for such a periodicity are not known at this time; further synoptic climatological studies of summer precipitation are warranted (cf. Barry, et al., 1982).

Summary

Analysis of long-term temperature and precipitation data for the Great Salt Lake drainage basin highlights the unusual nature of precipitation in recent years. Exceptionally large positive precipitation anomalies have been superimposed on a generally upward trend in seasonal precipitation. Fourteen of the last twenty water year totals have been above the long-term average and the years 1979-80, 1981-2, 1982-3 and 1983-4 were among the highest ever recorded. This accounts for the unprecedented rise in the level of the Great Salt Lake in recent years.

Acknowledgements

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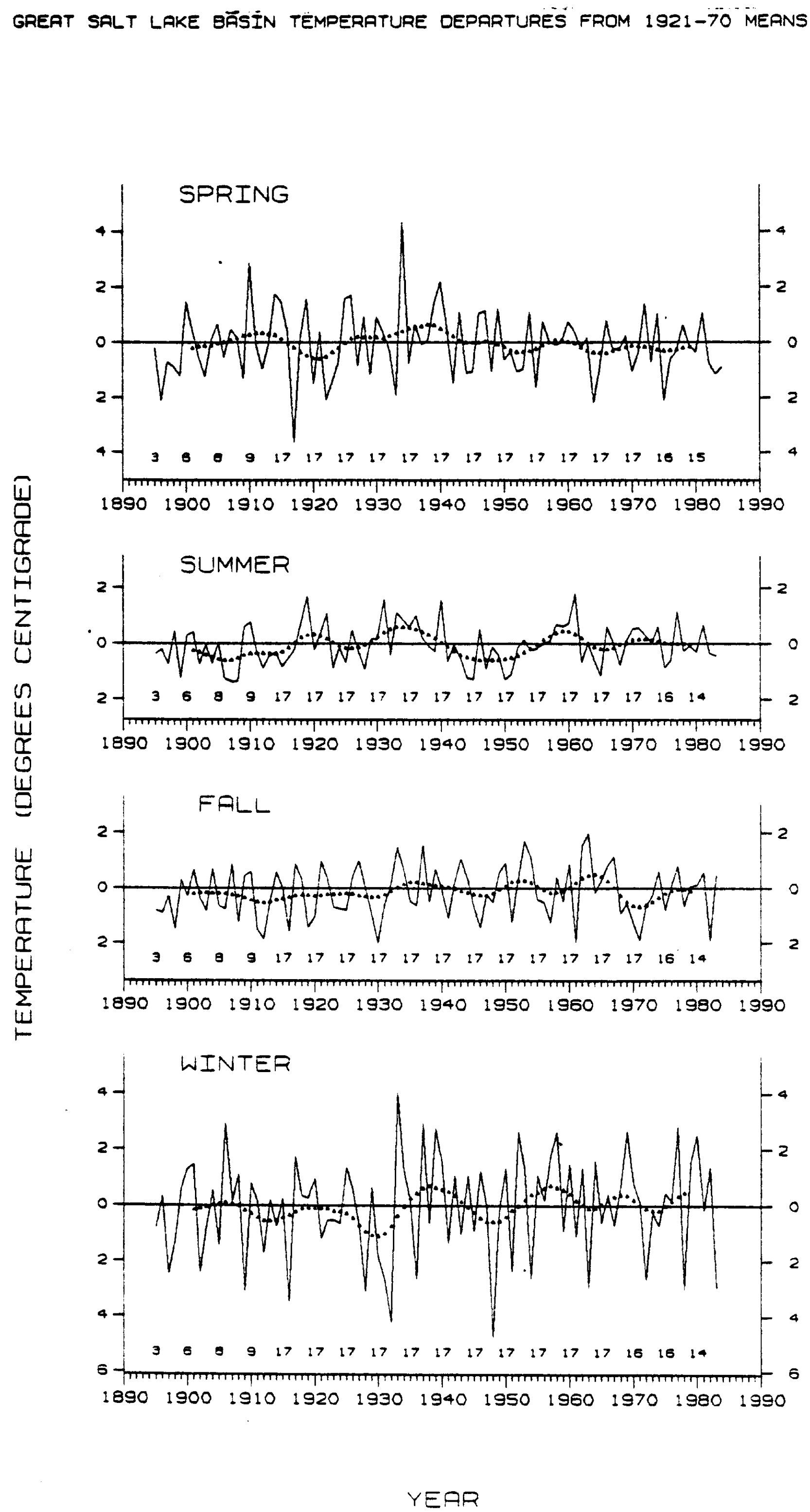

Figure 4. Seasonal mean temperatures expressed as departures from the 1921-70 period. Anomalies calculated for each station in Table 1b and averaged for the region (see caption to Figure 2).

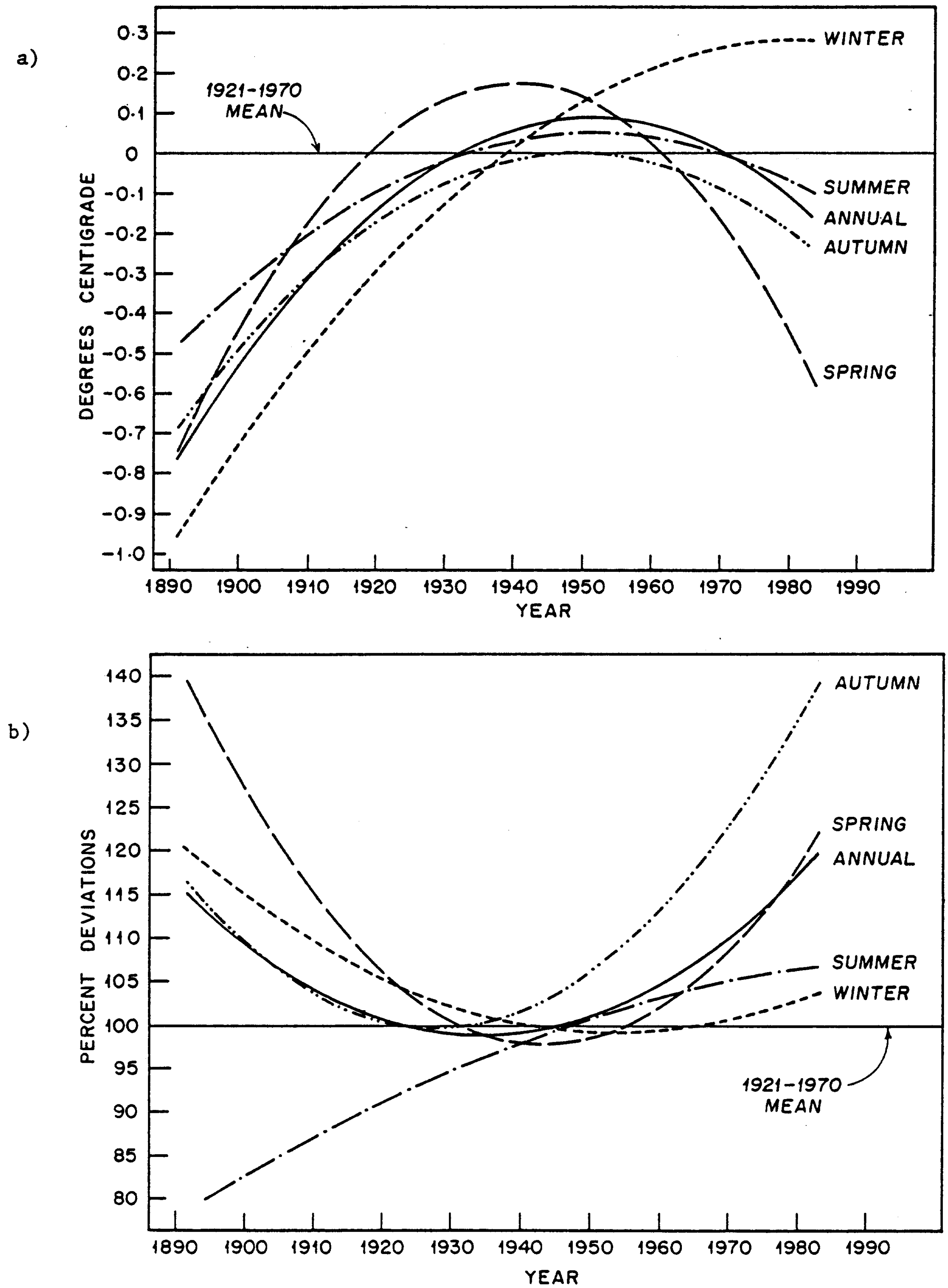

Figure 5. Generalized ("parabolic") trends in temperature (above) and precipitation (below) based on 3-point polynomial regression. Precipitation shows a general decline to around 1940 and an increase thereafter except in summer which has a general upward trend throughout the record. Temperature trends are generally the inverse (cf. Figures 2 and 4 and Table 3).

TABLE 3

Correlation Coefficients Between Temperature and Precipitation Data (1891-1983)

	r	*p*
Spring	-0.42	0.00002
Summer	-0.30	0.002
Fall	-0.38	0.00009
Winter	-0.03	0.39
Annual	-0.19	0.034

TABLE 4

Statistically Significant Periodicities (years)

a) Precipitation Data, 1891-1984

Spring	2.74
Summer	2.02
Fall	10.33
Winter	2.21
Annual	18.6

b) Temperature Data, 1891-1984

Spring	4.89
Summer	23.25
Fall	5.17
Winter	2.21
Annual	

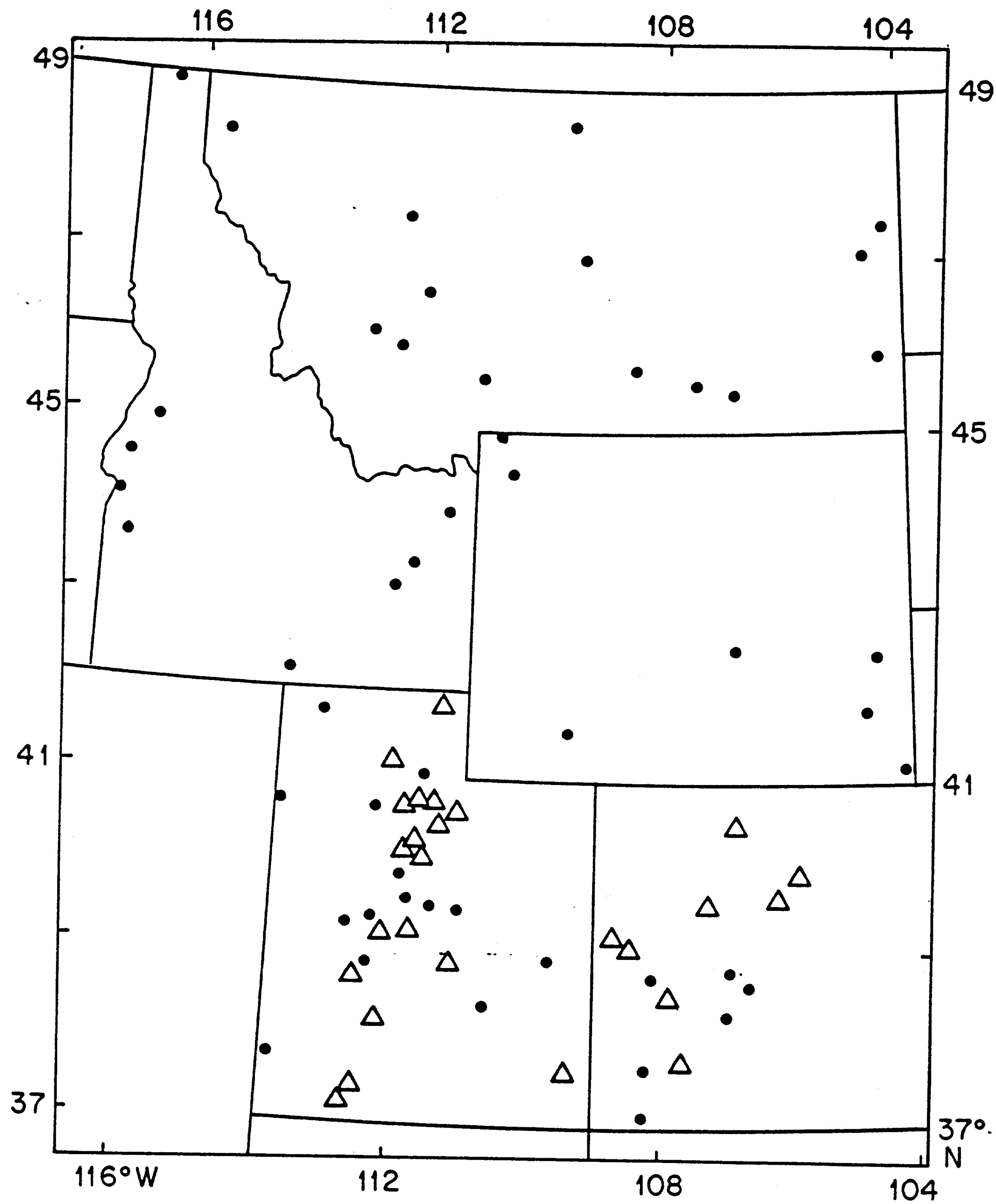

Figure 6. Distribution of long-term summer (J,J,A) precipitation records analyzed by power spectrum analysis. Those showing a statistically significant high frequency periodicity of 2 to 2.2 years are indicated by triangles and those showing no such periodicity by dots (from Bradley, 1976).

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Recurrent Patterns of Regional Streamflow Variability in the Western United States

Harry F. Lins U.S. Geological Survey

1. INTRODUCTION

Over the past four decades hydrologists have focused relatively little attention on synoptic or nationwide variations in annual streamflow. Instead, most research has been geared toward understanding the characteristics of shorter-term streamflow in individual basins or small regions. Of the broad-scale assessments which have been undertaken, most have emphasized data presentation and largely non-quantitative descriptive analyses (Harbeck and Langbein, 1949; Busby, 1963; Langbein and Slack, 1982). Very recently, however, several investigations have appeared in the literature which attempt to define, statistically, systematic spatial and temporal variations in streamflow across large parts of the United States (Bartlein, 1982; Meko and Stockton, 1984; Lins, 1985a; 1985b). These efforts may signal a significant conceptual shift in that hydrologists now appear to be looking at the hydrologic system in a more integrated way with the atmospheric system, not just in event space (as has long been done with rainfall-runoff models), but over climatic time scales. Such a shift is important since it is probable that reasonable long-term (seasonal to annual) forecasts of hydrologic conditions, if possible at all, will involve a coupling of climatic and hydrologic variables.

Given this perspective, the present study attempts to describe spatially coherent patterns of variation in annual streamflow over the western United States. Temporal variations in these patterns are identified and modeled in terms of their autoregressive-moving average (ARMA) properties. Finally, the relationship between these regional patterns of streamflow variability and annual changes in the level of the Great Salt Lake are identified.

2. DATA AND METHODS

Mean annual values of streamflow covering the western United States have been assembled for the period 1931-1978.[1/] The streamflow data, excerpted from Langbein and Slack (1982), were aggregated by 54 2.0° latitude-by-3.0° longitude grid cells (Fig. 1). The Langbein and Slack data were compiled using gaging station records primarily on watercourses reporting no regulation or diversion or where diversion amounted to less than 10 percent of the mean annual runoff. In a few instances, where station records could not be found meeting the above criteria, records including greater alterations due to regulation or diversion were used. In all such instances, preference was given to flow records least so affected, and the flows were adjusted to compensate for the alteration.

At most grid cells, positive skewness was observed in the frequency distribution of annual flows. A more symmetric distribution was obtained by plotting the frequency distribution of the logarithms of the flows. Normally distributed data were desired for subsequent principal component significance testing,

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1/ Throughout this report the terms "year" and "annual" refer to the water year which begins on October 1st and ends on September 30th of the following calendar year. Thus, water year 1931 began on October 1, 1930, and ended on September 30, 1931.

the statistic for which is derived from a multivariate normal population. Thus, the annual values were logarithmically transformed and then standardized. The standardization is given by

$$q'_{in} = \frac{q_{in} - \bar{q}_i}{s_i} \tag{1}$$

where q'_{in} is the standardized value of the log-transformed mean annual streamflow observation q_{in} at the i^{th} grid cell in the n^{th} year of the 48-year series; $\bar{q}_i$ is the 48-year mean value of q at the i^{th} grid cell, and s_i is the standard deviation of q from the 48-year mean at grid cell i.

From the 48x54 streamflow observation matrix systematic patterns of spatial variability in streamflow are represented in terms of the principal components (PC) of the matrices. Based on the recent literature documenting the efficacy of component rotation in deriving more physically meaningful patterns of variation (Karl and Koscielny, 1982; Walsh et al., 1982; Richman, 1983a; 1983b), an orthogonal (varimax) rotation is applied to the original unrotated components. In this regard, the unrotated components are derived quantities for which a basis exists to evaluate statistical significance. Current selection criteria for principal components have been designed for application to the results of unrotated solutions (Preisendorfer et al., 1981). Thus, if physically meaningful spatial patterns are sought, rotating only those components which first pass objective significance criteria provides a conservative framework for subsequent physical interpretation.

The percentage of variance and cumulative percentage of variance explained by the first five unrotated principal components of western U.S. streamflow appear in Table 1. The values reveal that the first five components of streamflow account for nearly 67 percent of the total variance. These components were tested for statistical significance using Preisendorfer and Barnett's (1977) dominant variance rule N. The rule N test is designed to determine if the eigenvalues, or percentages of variance, calculated in a principal component analysis (unrotated solution) of a geophysical data set are distinguishable from those produced from a spatially and temporally uncorrelated random process. As the rule is applied, the statistical significance of a component is suspect when its corresponding geophysical eigenvalue is less than that generated from a set of random data. Results of the application of rule N to the components of western streamflow appear in Table 2. As indicated in the table, the variances explained by the first four streamflow components exceed their random data counterparts. The variance explained by the fifth streamflow component (as well as those of all subsequent components), however, falls below its random data counterpart and thus represents the point at which the streamflow signal cannot be separated from noise.

3. SPATIAL VARIATIONS

The principal components of a spatial variable field can be plotted as contour maps and thus, in this example, provide a graphical depiction of the long-term variation in streamflow across the western United States. The values that are contoured on such maps are the component loadings, or the correlation coefficients between the streamflow variables and the principal components.

Loadings on the four orthogonally rotated components of streamflow are mapped in Fig. 2. The first component (PC-1-Or) is characterized by high loadings in the Pacific Northwest and Northern Rocky Mountain region. This pattern, accounting for 22.7 percent of the variance, emphasizes relatively strong departures from mean streamflow conditions in the Pacific Northwest, with a core of very high correlations in the immediate border area of Washington, Oregon, and Idaho. Another feature of this pattern is a zone of weaker, but negative, loadings across the Southwest indicating a contemporaneous above-/below-mean opposition in flows between the Pacific Northwest and the Southwest. This contrast has previously been noted by Lins (1985a) in a similar analysis of the entire United States, by Meko and Stockton (1984) using correlation analyses between water resource regions and subregions in the western U.S., as well as by Walsh et al. (1982) in the principal components of precipitation.

Loadings on the second component (PC-2-Or) indicate a region of interrelated streamflow departures stretching across California and the Great Basin to the Four Corners area. This pattern explains 16.0 percent of the variance. Component three (PC-3-Or) depicts regional streamflow anomalies in the Central Rockies and Central High Plains, while the fourth component (PC-4-Or) is characterized by flow anomalies centered in the Northern Great Plains. The variance accounted for by components three and four is 12.5 and 11.3 percent respectively. The spatial patterns described by PCs two and four also closely conform to regional modes of variation dominating annual streamflow across the conterminous United States described by Lins (1985a). Notably, while components one, two, and four collectively explain 50 percent of the variance in western U.S. streamflow, their counterparts in Lins' study explain nearly 33 percent of the total variance nationwide.

4. TEMPORAL VARIATIONS

The influence or importance of a principal component pattern through time is described in terms of its PC score. Component scores are a measure of the temporal similarity between the observed pattern of a variable in a given year and each principal component. They are computed here as the inner product between a streamflow observation and a principal component.

Scores for the four rotated components of streamflow appear in Fig. 3. Preliminary examination of these time series indicates that little or no persistence exists in the PCs of annual streamflow in the western U.S. This is not too surprising since raw annual flows, in general, have long been recognized as having relatively short memories. To understand better the nature of these principal component time series, autoregressive-moving average models are fit to the PC scores following the procedures of Box and Jenkins (1976).

Initially, persistence in the streamflow components is determined from the autocorrelation function (acf) plots in Fig. 4. It should be noted that the values of mean annual Western streamflow, which produced the acf in Fig. 4, were calculated as the average of the mean streamflow departures for the 54 grid cells for each of the 48 years of record. Neither mean annual, nor any of the first three components of flow exhibit persistence. Component four, the Northern Plains function, however, exhibits weak persistence and is the only flow component with a lag one correlation coefficient significantly different from zero.

Objective selection of model order for each streamflow series is made using Akaike's Information Criterion (AIC) (Akaike, 1974). Parameter estimates and AIC values for the models tested for each of the five process variables appear in Table 3. Scores for components one and two are best described as being independent. Scores for component three and mean annual streamflow most closely follow an ARMA (1,1) process. Component four scores are best characterized as fitting a high-order ARMA (3,3) model. A final, diagnostic check of the adequacy of each model was made by examining the acfs of the residuals to ensure that they appear as white noise. In all cases the residuals check confirmed the AIC selection.

6. RELATION TO GREAT SALT LAKE

Having described the broad regions of relatively homogeneous streamflow behavior across the Western United States, an attempt is made to determine what, if any, relationships exist between these patterns and variations in the level of the Great Salt Lake during the 48-year period of record. To do this, annual lake levels were first determined and are plotted in Fig. 5a. The values in the figure are lake levels observed at the end of each water year and thus represent an integrated intra-annual measure of input to, and losses from, the lake. Although it may not be easy to ascertain from Fig. 5a, an extension of the lake level curve back to 1857 indicates nonstationarity in the time series of Great Salt Lake levels. Thus, to avoid any bias due to trend, the first differences of the values in Fig. 5a were calculated for comparison to the PCs. A plot of the first differences appears in Fig. 5b.

Initially, cross-correlations between the annual differences (in Fig. 5b) and the annual mean, and each of the four principal components of streamflow (from Fig. 4) were determined. Results appear in Table 4. Positive correlations between all five streamflow variables and the differences in Salt Lake level are indicated. Highest correlations are with the Northern Plains function (PC-4-Or) and lowest with the Pacific Northwest (PC-1-Or). Significantly, a relatively high correlation also exists with mean annual streamflow. These findings are somewhat surprising since the highest PC loadings interpolated for the Great Salt Lake (~.50; see Fig. 2) are associated with PC-2-Or, the Far West-Great Basin function, and it was therefore expected that the highest correlation would be observed with this pattern. In fact, however, the loading values at Salt Lake associated with the other components are, for PC-1-Or .25; PC-3-Or .42; and PC-4-Or .46. Thus, little difference exists between PCs 2, 3, and 4, and the low loading associated with the first PC (.25) is consistent with the low correlation coefficient mentioned previously.

One possible explanation for the observed spatial distribution of correlations between Salt Lake levels and streamflow PCs may be associated with the overall strength of the westerlies. The high correlation with mean annual streamflow in conjunction with the uniformly positive correlations with all four principal components is suggestive of a synoptic-scale association with the general circulation. In general, annual streamflows in the four PC regions are most influenced by winter season (half-year) precipitation. So, too, are annual changes in the Great Salt lake. The intensification of the westerlies during the winter season produces monthly precipitation maxima in the regions of components one and two and snowfall in the higher elevations and latitudes. This latter point is significant in PC regions three and four where, although maximum precipitation generally occurs in summer months, flows are at a maximum in the latter months of the winter half-year as a result of snowmelt. Thus, winter season upper air patterns, hence precipitation, dictate in large measure the direction and magnitude of annual flows across the Western United States in general and annual changes in the level of the Great Salt Lake in particular. This finding is consistent with recent research specifically addressing relationships between precipitation in the western U.S. and upper air circulation (Cayan and Roads, 1984; Weare and Hoeschele, 1983; Klein, 1965).

7. SUMMARY

Four broad regions of the western United States within which annual streamflows exhibit strong spatial coherence have been identified using principal component analysis with a varimax rotation. These patterns are consistent with previously documented streamflow regimes as well as with upper air and precipitation modes of variation. Collectively, the four regional components account for nearly 63 percent of the total annual variation in western U.S. streamflow. Through time, scores on the first and second components most closely fit an independent process. Scores on the third component and mean annual streamflow indicate a first-order ARMA process, while those for component four best fit a third-order ARMA process. Weak persistence is observed in the scores of the fourth PC only. Correlation of the time-series of component scores with annual differences in the level of the Great Salt Lake indicates that the hydrology of the western United States, on an annual basis, is related to winter season synoptic climatology. Moderate correlations are indicated between annual changes in the level of Great Salt Lake and all regional flow components except the Pacific Northwest. It is important to note the close annual agreement between variations in the Great Salt Lake and in mean annual streamflow over the entire western U.S.

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Table 1.--Percentage of variance and cumulative percentage of variance explained by the first five principal components of streamflow in the Western United States

Principal Component	Percent Variance Explained	Cumulative Percent Variance Explained
1	27.33	27.33
2	20.29	47.62
3	8.41	56.03
4	6.55	62.58
5	4.05	66.63

Table 2.--Comparison of the streamflow percentages of variance explained with those generated by a random process

Principal Component	Percent Variance Explained		T_j/U_j^{95} 2/
	Streamflow Data (T_j)	Random Data (U_j^{95}) 1/	
1	27.33	8.48	3.22
2	20.29	7.53	2.69
3	8.41	6.81	1.23
4	6.55	6.33	1.03
5	4.05	5.85	0.69

1/ Computed for a 48x54 matrix.

2/ If $T_j/U_j^{95} \geq 1.00$, pass rule N; if < 1.00, fail.

Table 3.--Fitted model parameters for mean annual, and the first four principal components of mean annual, streamflow

Variable	Model	Parameter Estimates	AIC
Mean Annual	Independent	---	-80.801
Streamflow	(1,0)	$\hat{\phi}_1$ = -0.018	-78.817
	(1,1)*	$\hat{\phi}_1$ = 0.819	-81.019
		$\hat{\theta}_1$ = 0.956	
	(2,1)	$\hat{\phi}_1$ = -0.073	-76.039
		$\hat{\phi}_2$ = -0.190	
		$\hat{\theta}_1$ = -0.078	
PC-1-Or	Independent*	---	123.686
	(1,0)	$\hat{\phi}_1$ = 0.019	125.669
	(1,1)	$\hat{\phi}_1$ = 0.912	125.181
		$\hat{\theta}_1$ = 0.804	
	(2,1)	$\hat{\phi}_1$ = 0.661	125.679
		$\hat{\phi}_2$ = 0.204	
		$\hat{\theta}_1$ = 0.693	
PC-2-Or	Independent*	---	112.337
	(1,0)	$\hat{\phi}_1$ = 0.018	114.323
	(1,1)	$\hat{\phi}_1$ = -0.224	116.314
		$\hat{\theta}_1$ = -0.246	
	(2,1)	$\hat{\phi}_1$ = -0.885	117.352
		$\hat{\phi}_2$ = -0.045	
		$\hat{\theta}_1$ = -0.926	
PC-3-Or	Independent	---	102.939
	(1,0)	$\hat{\phi}_1$ = 0.047	104.834
	(1,1)*	$\hat{\phi}_1$ = 0.704	100.726
		$\hat{\theta}_1$ = -0.955	
	(2,1)	$\hat{\phi}_1$ = 0.944	105.719
		$\hat{\phi}_2$ = -0.146	
		$\hat{\theta}_1$ = -0.951	
PC-4-Or	Independent	---	104.834
	(1,0)	$\hat{\phi}_1$ = 0.388	99.220
	(1,1)	$\hat{\phi}_1$ = 0.914	96.990
		$\hat{\theta}_1$ = 0.648	
	(2,1)	$\hat{\phi}_1$ = 0.913	98.990
		$\hat{\phi}_2$ = 0.001	
		$\hat{\theta}_1$ = 0.647	
	(3,1)	$\hat{\phi}_1$ = 0.803	100.066
		$\hat{\phi}_2$ = -0.087	
		$\hat{\phi}_3$ = 0.177	
		$\hat{\theta}_1$ = 0.528	
	(3,3)*	$\hat{\phi}_1$ = 1.088	93.051
		$\hat{\phi}_2$ = 0.460	
		$\hat{\phi}^3$ = -0.662	
		$\hat{\theta}_1$ = 0.935	
		$\hat{\theta}_2$ = 0.707	
		$\hat{\theta}_3$ = -0.935	
	(4,4)	$\hat{\phi}_1$ = -0.025	103.225
		$\hat{\phi}_2$ = 0.027	
		$\hat{\phi}_3$ = 0.340	
		$\hat{\phi}_4$ = 0.393	
		$\hat{\theta}_1$ = -0.408	
		$\hat{\theta}_2$ = -0.290	
		$\hat{\theta}_3$ = 0.010	
		$\hat{\theta}_4$ = 0.614	

* Denotes 'best-fit' model based on Akaike's Information Criterion (AIC).

Table 4.--Cross-correlation coefficients (lag k=0) between first differences in annual Great Salt Lake levels and the mean and principal components of annual streamflow in the Western United States. A zero-lag correlation of 0.29 is significant at the 0.05 level.

	Correlation (r) with First Differences in Great Salt Lake Level
Mean Annual Streamflow	0.64
PC-1-Or (Pacific Northwest)	0.33
PC-2-Or (Far West-Great Basin)	0.58
PC-3-Or (Rockies-High Plains)	0.53
PC-4-Or (Northern Plains)	0.74

Figure 1.--Centroids of the 54 2.0 latitude-by-3.0 longitude grid cells containing streamflow data used in the principal component analysis.

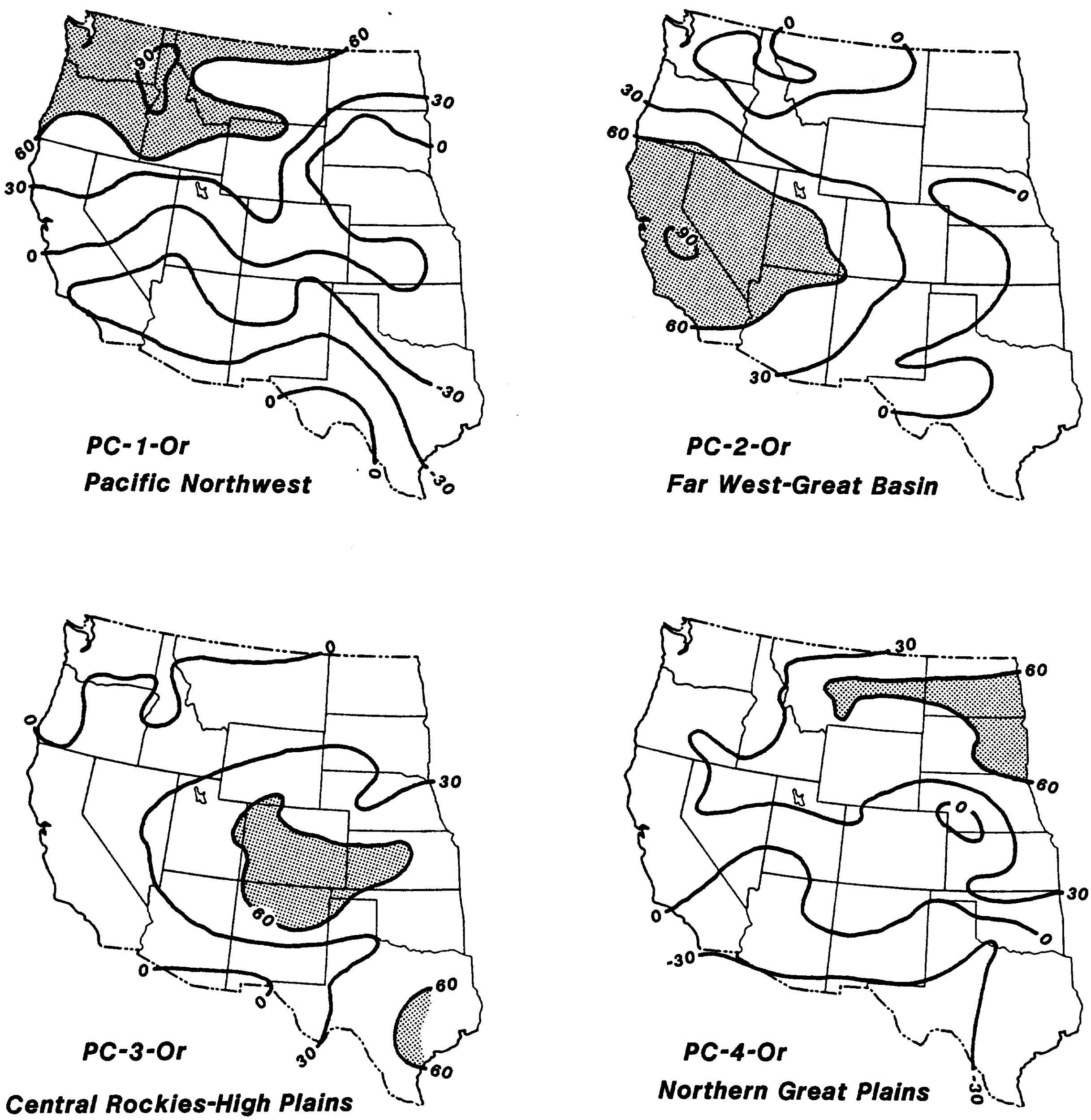


Figure 2.--Loadings (x 100) on the four orthogonally rotated principal components of annual streamflow.

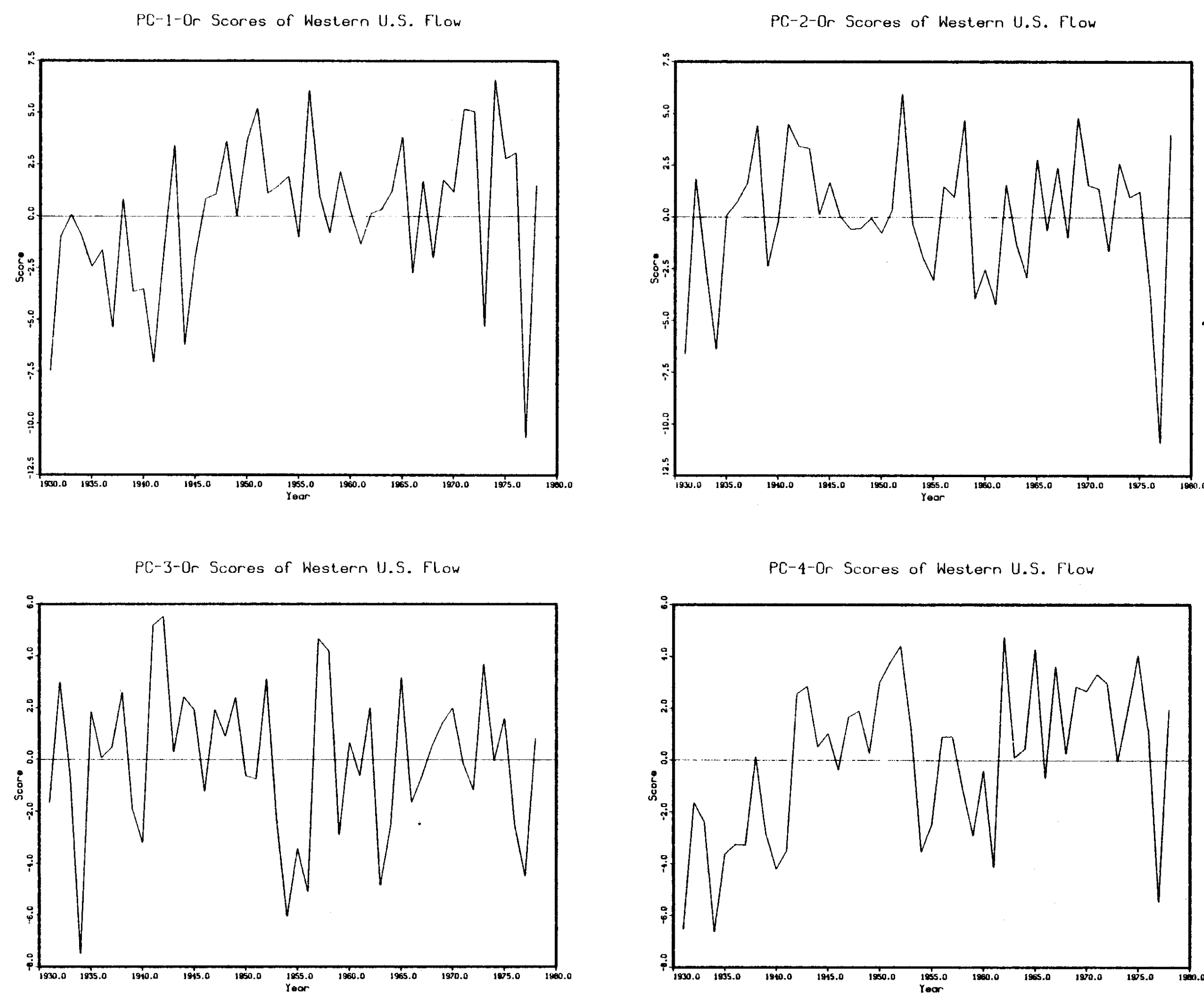


Figure 3.--Temporal variation in the annual scores (dimensionless) of the four orthogonally rotated components of annual streamflow.

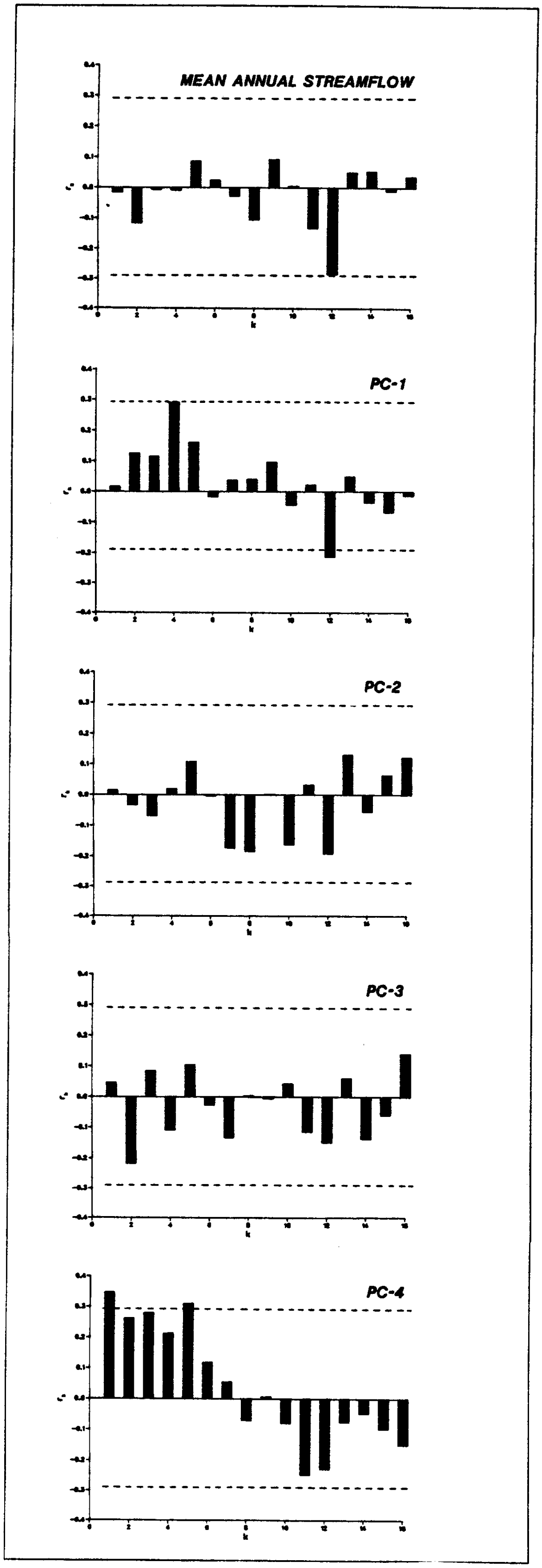
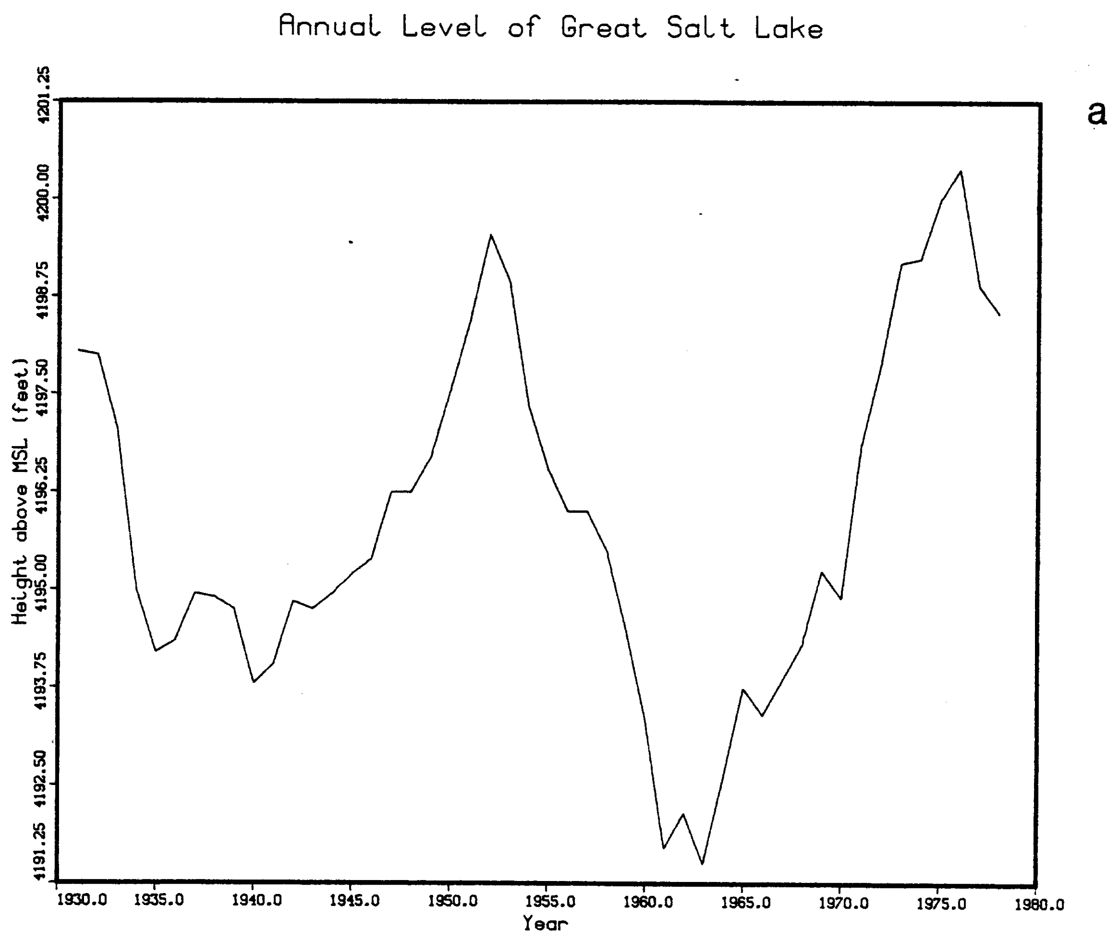


Figure 4.--Autocorrelation functions for mean annual and the four principal components of western U.S. streamflow. Dashed line represents the 0.05 significance level given by $\pm$ 1.96 $(1/(n-3)^{\frac{1}{2}})$ - 0.292.



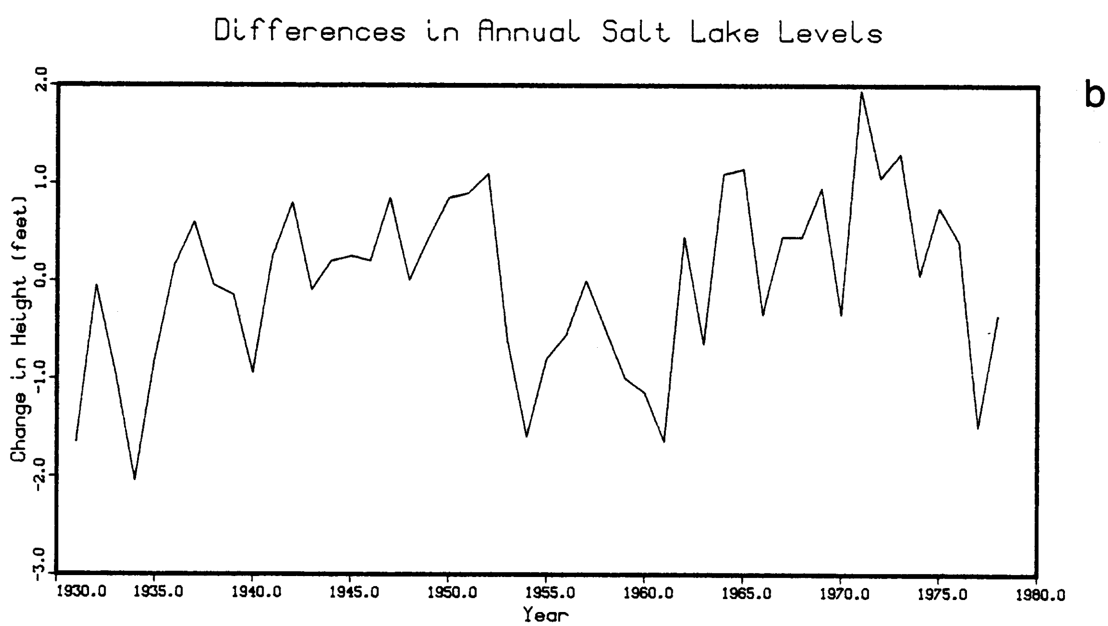


Figure 5.--Annual level (in feet above mean sea level) of the Great Salt Lake (a); and first differences (in feet) in those levels (b).

Paul A. Kay and Henry F. Diaz

1. INTRODUCTION

As a closed lake in a basin of interior drainage, the Great Salt Lake (GSL) (Fig. 1) is sensitive to climatic variability. The volume of water in GSL represents a balance between input, as surface and groundwater flows from basin and as precipitation directly onto the lake, and output as evaporation. The surface elevation of the lake is related, non-linearly, to lake volume, so that with allowances for lags in response, lake level fluctuations are reflections of the precipitation and evaporation climatology. In this paper, we examine monthly and seasonal climatic variables to find those most closely related to lake level behavior in recent decades.

The record of annual lake levels, either as annual peak or at the end of the water year, exhibits a very strong persistence with the first order serial correlation being about 0.98 (James *et al.* 1979). The annual level represents a balance between seasonal accumulation and loss, each having its own climatic characteristics, and therefore this single variable may mask some of the direct relationships. For these two reasons, we have decomposed the lake level record into two components, the seasonal rise and the seasonal fall. The lake typically peaks in May or June, and experiences its minimum stand in August or September, although earlier and later occurrences are recorded for both maximum and minimum (Arnow 1984). In this paper the seasonal rise for year t is calculated as the maximum in year t less the minimum in year t-1, and the seasonal fall for year t is calculated as the minimum in year t less the maximum occurring earlier in the same year. The two seasonal series (Fig. 2) exhibit considerably less persistence than the annual series; first order serial correlations are .20 for the rise series and .11 for the fall series, neither value being significant at the .05 level. As the modelling in this paper utilizes regression analyses, it is appropriate that autocorrelation be absent from the variables so that the assumptions of the technique are satisfied (Cliff and Ord, 1981).

We hypothesize that seasonal rise of GSL is primarily a function of accumulation season precipitation, and that seasonal fall is primarily a function of summer temperature (a surrogate for evaporation).

2. DATA

Lake level data are reported monthly by the U.S. Geological Survey in Water Supply Papers. The lake basin has experienced considerable human impact, in the form of water management and withdrawals. Reconstructed pristine levels (or natural levels), estimated as if there had been no human impact, were calculated up to 1968 (Whitaker, 1971). Since that time, it is estimated that no additional impacts were manifest, so that an adjustment of about 5+ feet representing the difference between actual and natural levels has been applied to the record to date (Arnow, personal communication, 1985).

Monthly mean temperatures and precipitation for the North Central (NCD) and the Northern Mountains (NMD) climatic divisions of Utah were obtained from NOAA's National Climatic Data Center in Asheville, N. C. for 1931 to 1984. These two divisions encompass the area which is the source of most of the inflow of water to Great Salt Lake (Fig. 1; cf. Arnow, 1984, Fig. 8). However, the divisions do not encompass all of the GSL basin, and NMD also comprises considerable area outside of the basin. Although we can expect some misspecification in the modelling work that follows, the number of stations in NMD outside the GSL basin is small. Therefore, the bias should also be minimal.

The climatic data were composited into a number of different seasons, depending on the needs of the models. Besides monthly data, we also use seasonal means or sums defined in one instance as the standard four meteorological seasons (SON = autumn, DJF = winter, MAM = spring, JJA = summer), and in another as the accumulation

(September to May) and loss (JJA) seasons. The divisions are used separately in one analysis, while in another the precipitation amounts in the two divisions are combined to represent total input.

Additionally, we examined monthly data for percent possible sunshine at Salt Lake City. We anticipated that winter precipitation would be negatively correlated and that summer evaporation would be positively correlated with these variables.

3. METHODS

The interrelationships between climatic variables within and between climatic divisions are investigated with Pearson correlation coefficients, tested as the Models of seasonal control on lake level rise and fall are created by multiple regression analyses. Since the "independent" variables in the models are interrelated, it is necessary to control for multicolinearity which might bias the regression results, leading to an inefficient and misspecified model. In our work, we have chosen the method of all possible subsets regression to examine many likely candidates for "best" model. Our selection criterion is Mallows' Cp statistic (Draper & Smith, 1981). Mallows' Cp measures the ratio of model reduction of variance to that possible in the theoretically correct model (approximated by the full model containing all predictor variables), traded off against the number of variables in the model (some of which might be spurious).

4. SECULAR PRECIPITATION FLUCTUATIONS IN THE VICINITY OF GREAT SALT LAKE

Before describing the interannual and longer-term variability of precipitation around the GSL, we should first examine the annual cycle of precipitation in the GSL region. As noted earlier, the variation of precipitation associated with the seasons is the major factor controlling the long-term "equilibrium" level of GSL and the variability around the mean annual curve is primarily responsible for its year-to-year fluctuations.

Figure 3 gives the annual cycle of precipitation in the two climatic divisions surrounding the GSL. Both show a minimum in summer and a maximum in winter and spring. The NCD, which is the lower elevation division, exhibits a maximum during the spring months of March-May, while the NMD exhibits a more uniform precipitation maximum that extends from December-April.

The average variability about the mean is given in terms of the coefficient of variability (CV = $(\sigma/\bar{x}) \times 100$), where σ is the standard deviation about the long-term mean, $\bar{x}$. In both divisions, September exhibits the greatest relative year-to-year variability, and both show a secondary variance maximum during June. For the NCD, in fact, the standard deviation for September is the highest of any calendar month. Interannual variability relative to the mean is minimum during the wet season from about November-April.

The next five graphs show cumulative departure curves for five different seasonal groupings. Cumulative departures (CD) from the mean are a useful tool for helping to identify periods of sustained anomalies.

The first graph (Fig. 4) shows CD for the period September-May and therefore excludes the driest season which is summer (June-August). It is clear that the period 1981-82 to 1983-84, particularly the last two of these three years were exceptionally wet in comparison with the previous 4-5 decades. The next graph (Fig. 5) gives the CD curve for the October-April season, which represents the wettest part of the annual cycle. This curve is similar to the one in Fig. 4 but shows a comparatively smaller recent rise. A similar but longer-lived wet regime occurred during the 1940s.

Figure 6 illustrates the fluctuations in the water-year total (Oct.-Sept.), and again shows an unprecedented rise for water years 1981-82 through 1983-84. By contrast, the winter months of December-February (Fig. 7) have a different mix of wet and dry seasons. Winter precipitation during the past few years has been only slightly above normal.

In some ways the most interesting curve is the summer CD trace shown in Fig. 8. This appears to show a somewhat regular rhythm of dry and wet spells occurring about 20 years apart. Fig. 9a shows the variance spectrum of this curve which indeed indicates a statistically significant peak in the neighborhood of 20 years. However, because a cumulative departure curve is effectively a low-pass filtered series, it cannot tell us anything about the high-frequency end of the precipitation spectrum. Fig. 9b gives the variance spectrum of the individual summer season anomalies and this shows, in addition to the ~ 20-year oscillation a significant concentration of variance around 2-2.2 years which is the range of the quasi-biennial oscillation (QBO).

Before we take a further look at the nature of these quasi-periodic fluctuations, it is worthwhile noting that as a whole, the spell of wet weather that began around September 1982 is quite anomalous based on the available period of record. Fig. 10 shows the CD curve of the monthly values from January 1931 to September 1984. Only the unusually dry spell that occurred during the 1930s is close to matching (although in the negative sense) the steep rise of the recent years. The statistical significance of this wet spell is illustrated in Table 2 which gives the probability of obtaining a sample mean of 25 months in length (September 1982-September 1984) as wet, or as different from the mean as that which was observed if it was drawn at random from the same population. It is clear that this is an unusual occurrence with a recurrence probability of something less than once in 200 years.

5. INTERRELATIONSHIPS BETWEEN CLIMATIC VARIABLES

The annual behavior of the correlations between monthly temperature and precipitation within each climatic division is very similar (Fig. 11). The correlations are negative in all months except January and February (only February in the Northern Mountains), but are significant (at the .05 level) only in Sept.-Nov. and April-June. The correlations are strongest in June. These relationships are consistent with the correlations using seasonal data given in Table 1.

We would expect there are very strong positive correlations between temperatures and between precipitation in the two divisions (Fig. 12 and Table 1). The correlations between divisional temperatures are all above .9 except in July and August, when perhaps a higher frequency of cloudiness in the mountains interferes with the characteristic temperature of the air masses. The correlations between divisional precipitation are lowest in November to January, perhaps because of differential orographic influences that operate at these times. Relatively lower correlations in July and August probably reflect the variable spatial nature of summer thunderstorm precipitation and orographic effects.

The percent of possible sunshine correlates with temperature and precipitation in both divisions in a manner consistent with the variable influence of cloudiness on temperature and precipitation during the year (Fig. 13). Precipitation is negatively and significantly correlated with percent possible sunshine in all months. The relationship is strongest in spring and autumn months. Temperature is significantly positively correlated with percent possible sunshine only in October and April to June; these are the same months that have the strongest sunshine-precipitation correlation. A negative correlation in winter, although not significant, is consistent with the relationship expected for cold dry air masses.

6. MODELS OF SEASONAL CONTROL

Accumulation and loss seasons

Seasonal lake level rise is best modeled by accumulation season (September–May) precipitation in both divisions and by summer (June–August) precipitation and summer sunshine in NCD. These four variables account for 78% of the variance of lake-level rise, and yield estimates with a standard error of ± 0.49 ft. of rise (table 4). The addition of the previous year's rise adds only about 1.5% to R^2 and the coefficient is not significant. The model makes physical sense, except for the positive relationship to percent possible sunshine in spring. We would expect an inverse relationship, especially as the March to April precipitation is in the model with a positive coefficient. We can only conclude that the inclusion of the sunshine variable is an artificial effect due to multicolinearity among the variables. These results tend to confirm the hypothesis regarding the central importance of accumulation season precipitation for lake level rise.

Seasonal lake level fall is best predicted by a model containing summer precipitation in NCD and accumulation season precipitation in NMD (Table 5). The model accounts for 78% of the variance in lake level fall, and predicts lake level fall with a standard error of estimate of ± 0.48 ft. The relationship to both precipitation variables is negative. It is surprising that temperature does not appear as a significant predictor in this, or any of the lesser candidates for "best", model. Summer precipitation around the lake appears to be an important control. Winter precipitation also is important, probably through its relationship to lake level rise, with which lake level fall is strongly and negatively correlated (that is, the stronger the preceding lake level rise, the smaller the subsequent fall). The hypothesis regarding the central importance of summer temperature for lake level fall must be modified to account for the rate of summer precipitation.

Meteorological seasons

Seasonal lake rise is best predicted by a model containing mostly precipitation variables (Table 6). Autumn, winter and spring precipitation in the NCD, and winter precipitation in the NMD, are most important. Cool spring temperatures and sunshine in the NCD are also significant predictors. The lake level rise in the previous year enters the equation, but its coefficient is not significant and it adds little to the R^2. This model is similar to that of the preceding section. Here, the model accounts for 82% of the variance in rise, with a standard error of estimate of ± 0.45 ft of lake level rise.

Seasonal lake level fall is best predicted by a model containing precipitation variables (Table 7). Antecedent precipitation, and summer precipitation in the NCD are most important. The relationships are negative. Temperature does not enter the equation as a significant predictor. The model accounts for 81% of the variance in lake level fall, with a standard error of estimate of ± 0.46 ft of lake level fall.

Model-Tests of Hypothesis

Because of the intercorrelations of precipitation between the two divisions, it was decided to sum the precipitation values to represent total inflow to the lake. Temperature for NCD only was used, as the lake lies mostly in that division and not at all in the Northern Mountains. Sunshine was omitted, both because of its intercorrelation with temperature and precipitation and because of problems noted above.

Accordingly, lake level rise was regressed against the combined precipitation for autumn, winter and spring, and the NCD temperature for these three seasons. The best model included the three precipitation variables, and spring temperature (Table 8). The model accounted for 73% of the variance in lake level rise, with a standard error of estimate of ± 0.43 feet. The residuals from the model were approximately normally distributed and were free from significant autocorrelation, indicating no bias in the model. The hypothesis appears to be substantiated.

The order of importance of the predictor variables is of some interest. Autumn precipitation is the most important variable, followed by winter then spring precipitation. It appears that early season accumulation is most effective in setting the stage for a large rise of the lake, perhaps by saturating the ground and establishing the snowpack.

The seasonal lake level fall was regressed against NCD temperature and combined precipitation for spring and summer seasons. The best model included the precipitation variables, and the rise of the lake in the preceding accumulation season (Table 9). All predictors were inversely related to fall; that is, the magnitude of the fall would be largest when precipitation and the preceding rise were smallest. The model accounted for 68% of the variance in fall, with a standard error of estimate of $\pm$ 0.25 feet. Summer precipitation was the most important variable, with preceding rise next. Without rise as a predictor, a model with spring and summer precipitation and summer temperature was produced, but the coefficient for summer temperature (although positive) was not significant and the overall performance of the model was worse. Thus it appears that precipitation, rather than temperature is the most important control of the magnitude of lake level fall in summer. Furthermore, the inclusion of all five independent variables into the model produces an R^2 of only 68% which is the same as the 3-variable model described above.

7. CONCLUSIONS

Our main findings can be summarized as follows. Lake level fluctuations and basin wide precipitation exhibit a great deal of parallelism at time scales greater than one month (the shortest time resolution for climate variable used in this study) up to the order of several decades (the available length of record).

In establishing statistical associations between climatic variability and the lake level record, it is important to treat the lake level rise and fall separately. In a broad sense, lake level rise is a function of accumulation season (Sept–May) precipitation and secondarily to summer precipitation and summer sunshine (temperature does not enter explicitly into the regression model). Lake level fall is principally a function of summer precipitation in the lake's environs (the North Central Division) and accumulation season precipitation in the snowpack source region of the NMD. Once again the contribution of temperature to the variance of lake level fall is not statistically significant.

A finer resolution model using standard 3-month climatological seasons is almost as good as the "two-season" model in terms of R^2, but provides a great deal more insight into climate-lake level interaction. This model shows that lake level rise is best predicted (in order of importance) by autumn, winter and spring precipitation in NCD and winter precipitation in NMD. Combining the precipitation totals for the two divisions results in a model with the same seasonal predictors, i.e. autumn, followed by winter and spring precipitation with spring temperature in NCD entering the regression fourth. Lake level fall in the combined seasonal model shows summer precipitation to be the single most important prediction followed by previous seasonal lake level rise and by spring precipitation.

There are two important time scales of variability evident in the precipitation record that have a significant impact on lake level fluctuations. The longer time scale is on the order of 20 years. It is most clearly defined during the summer season where it has a peak to trough amplitude of $\sim$ 4 in. The second principal scale of variability is in the neighborhood of 2 years, and has an amplitude of $\sim$ 3 in.

Based on the record of the past 50 years, it appears likely that the lake level will again trend downward over the next 5–10 years, although exactly when this trend will set in is presently unclear.

8. REFERENCES

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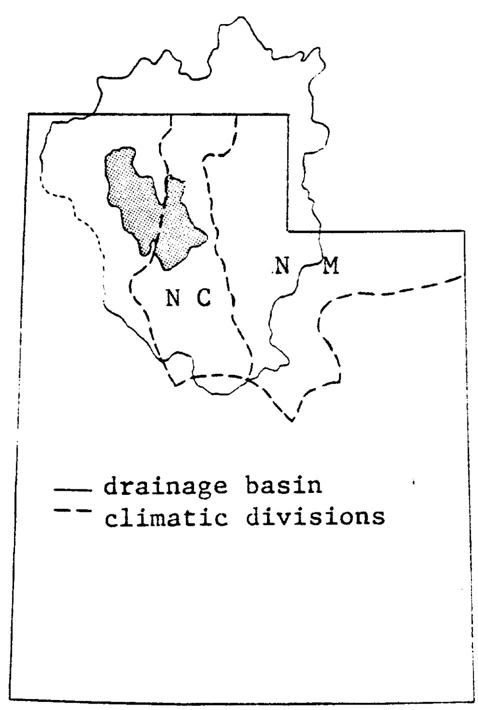


Figure 1. Location of Great Salt Lake, its drainage basin, and the North Central (NC) and Northern Mountains (NM) climatic divisions.

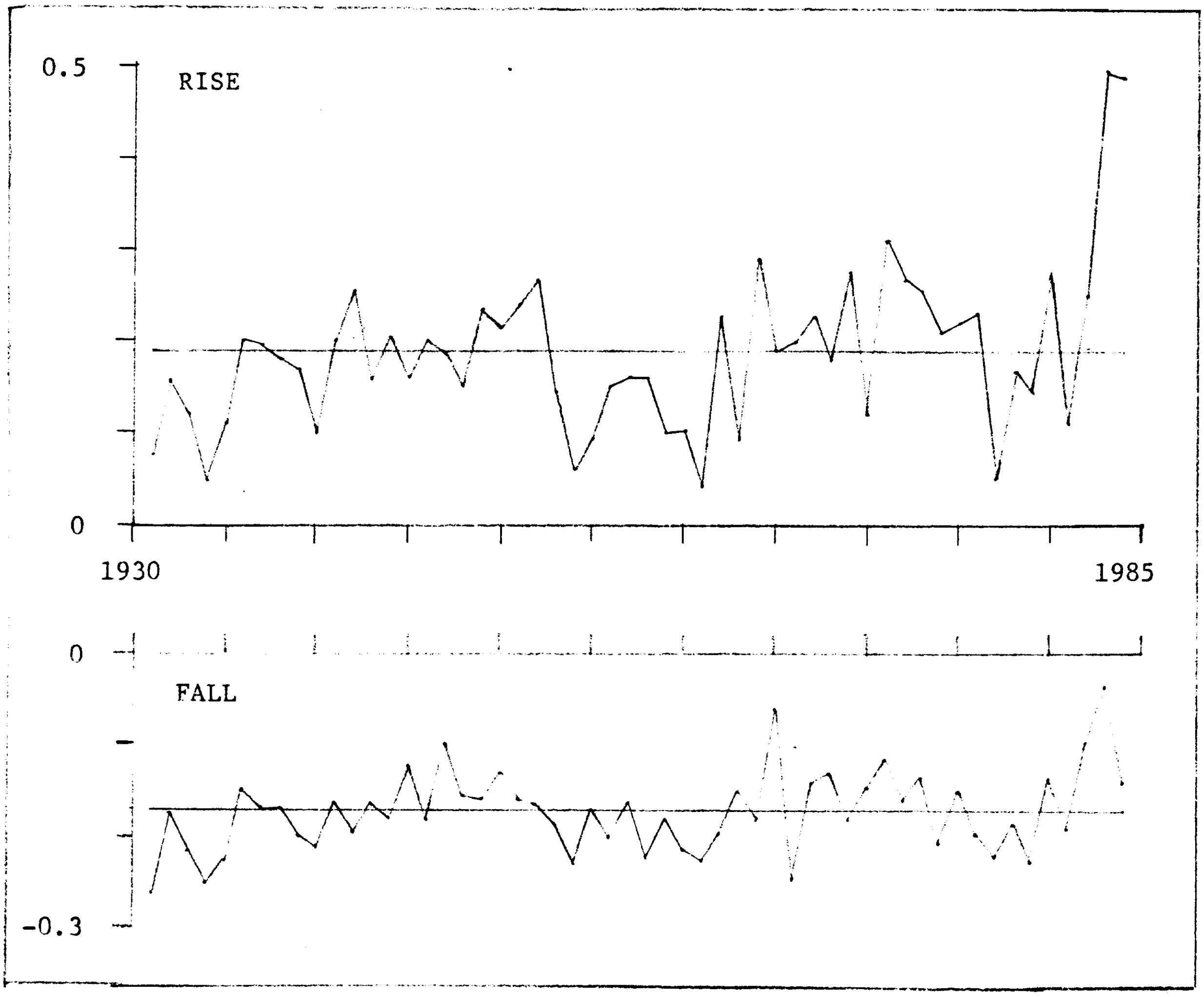


Figure 2. Annual values of lake rise (annual maximum minus previous year's minimum), top curve, and of lake fall (annual minimum minus annual maximum), bottom curve. Scale is in feet.

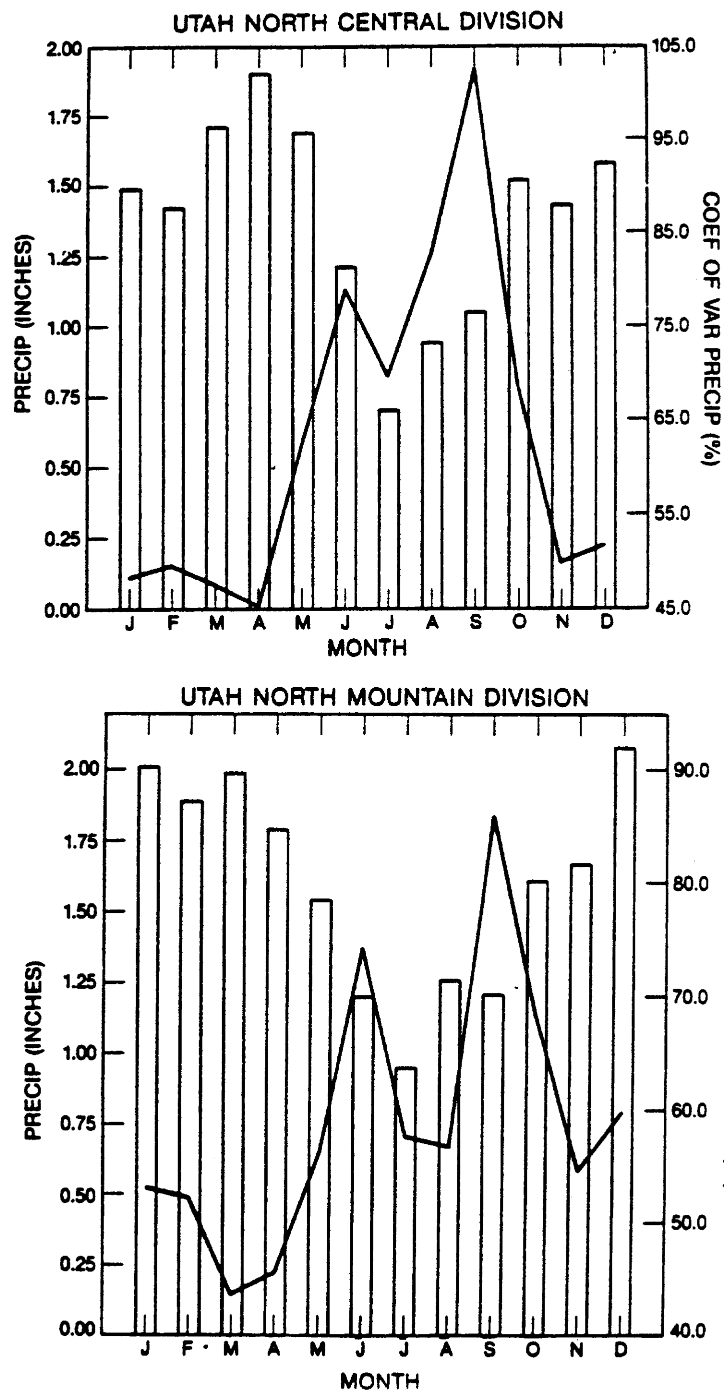


Figure 3 Annual cycle of precipitation and precipitation variability for the North Central and North Mountain climate divisions. Variability is given in terms of the coefficient of variation (see text).

Figure 4 Cumulative departure curve of September-May (accumulation season) aggregate precipitation totals for both climate divisions.

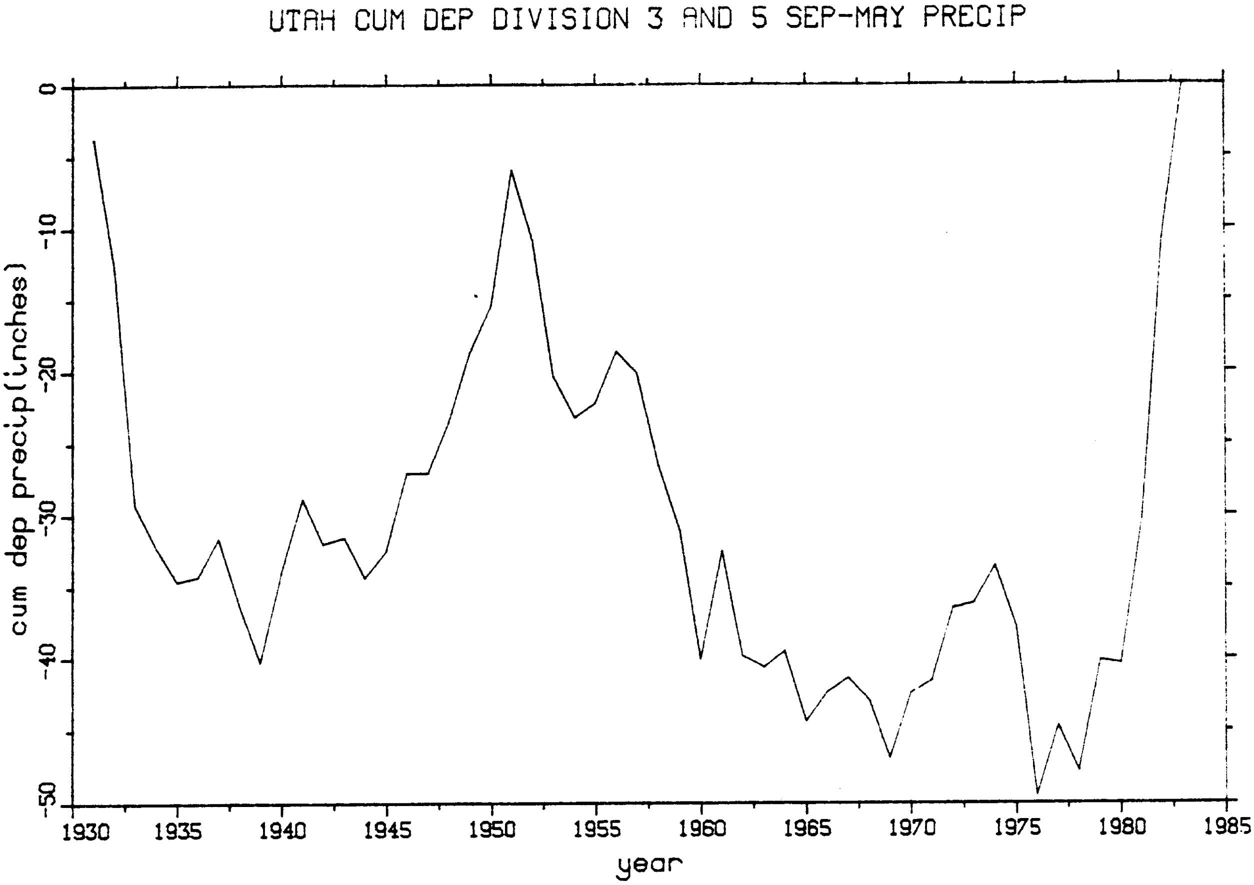


Figure 5 As in Fig. 4, except for October-April season corresponding to the wettest part of the water year.

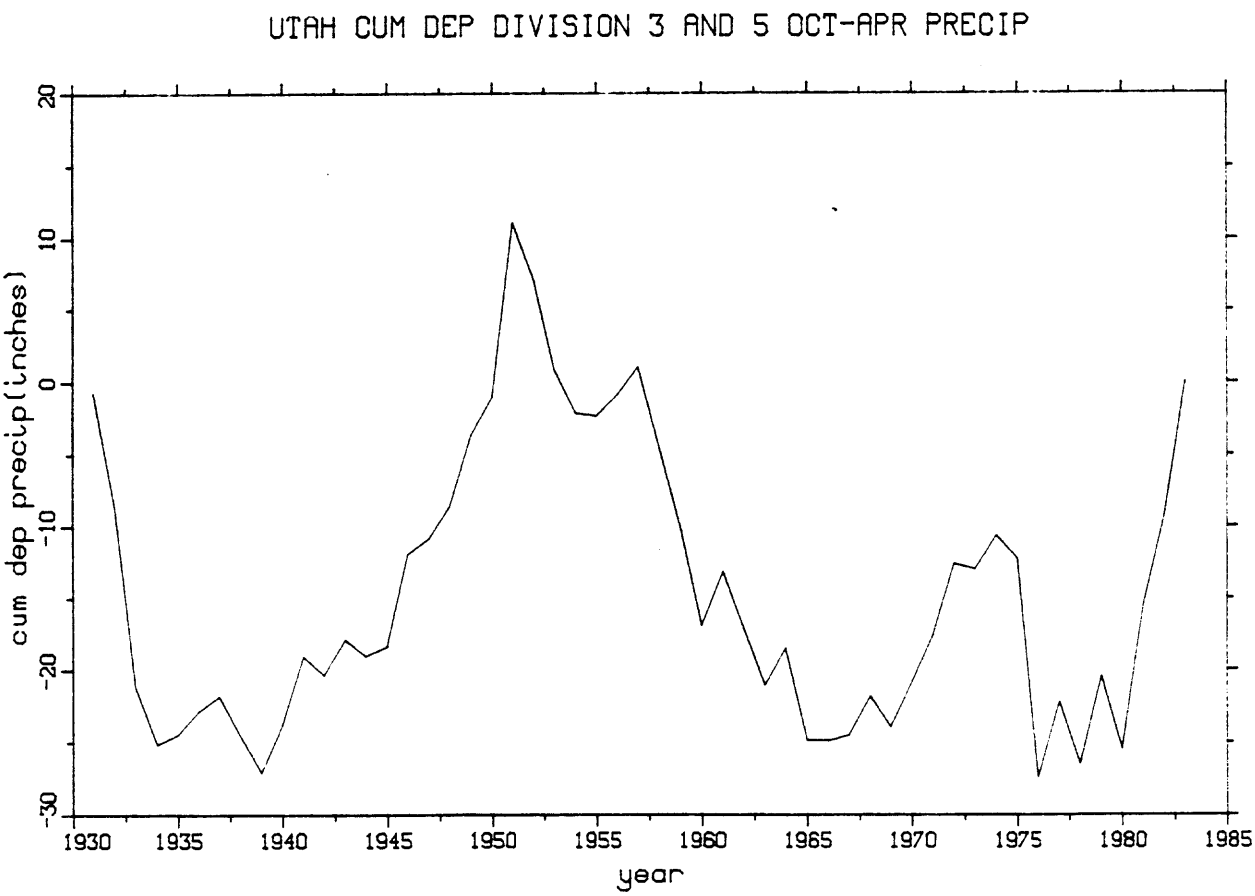


Figure 6 As in Fig. 4, except for the water year of October-September.

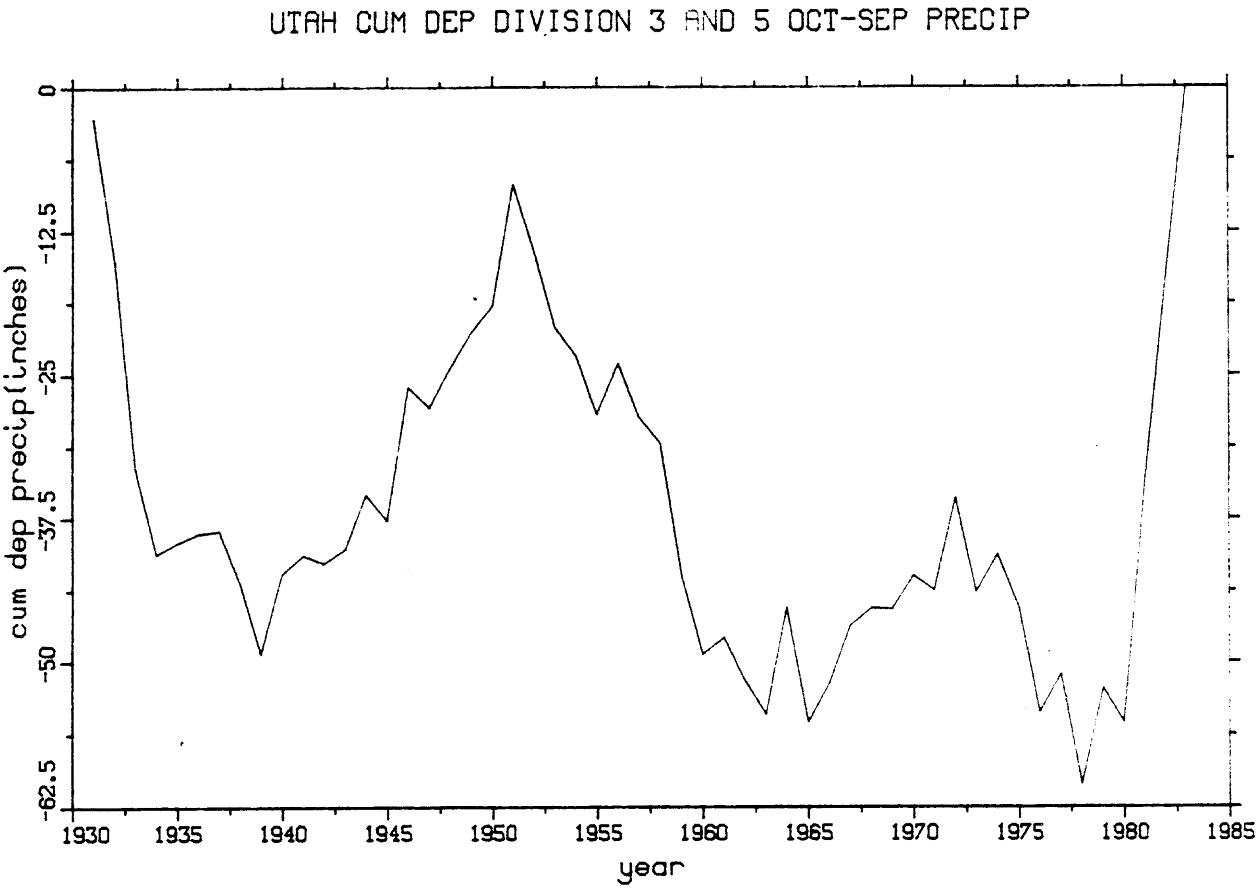


Figure 7 As in Fig. 4, except for the winter months of December-February.

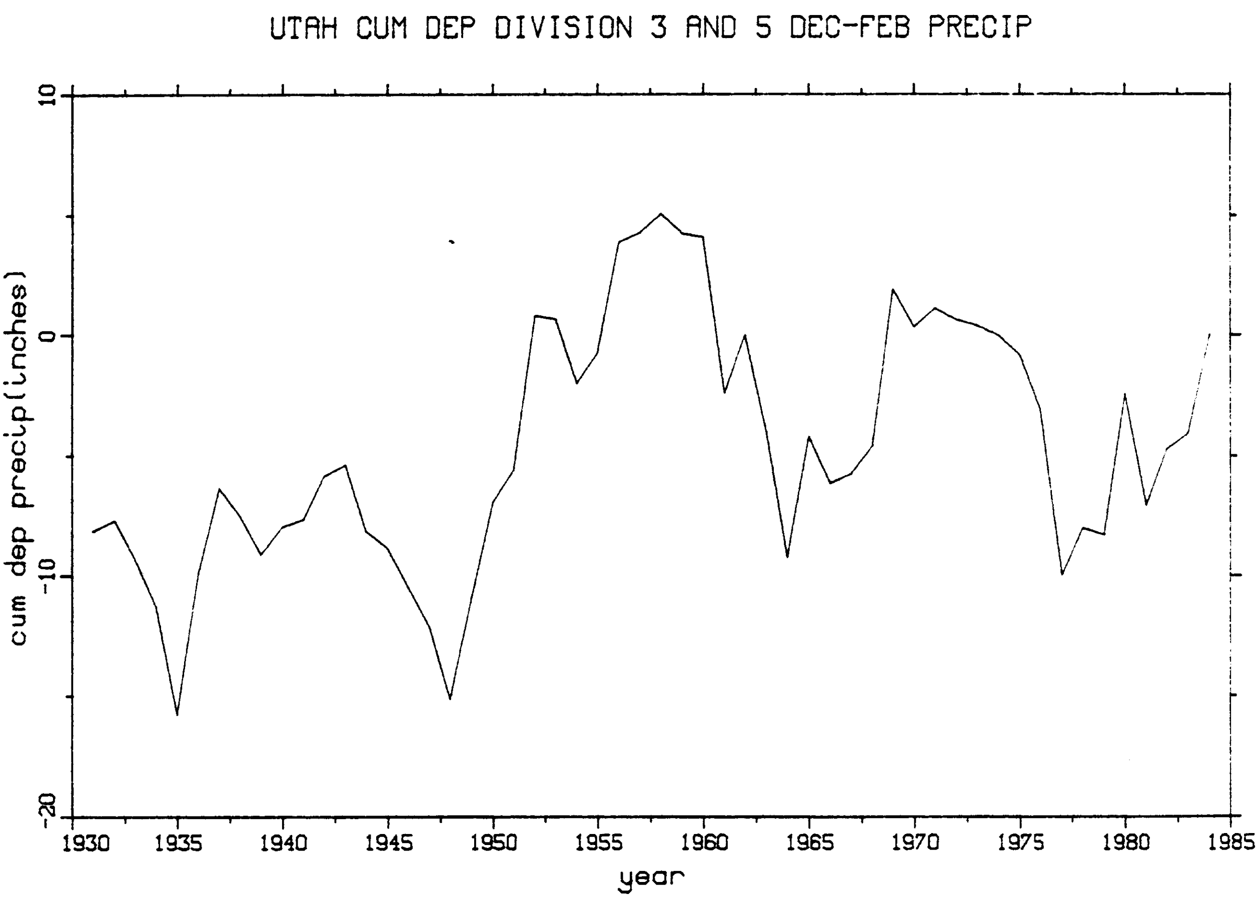


Figure 8 As in Fig. 4, except for the summer months of June-August.

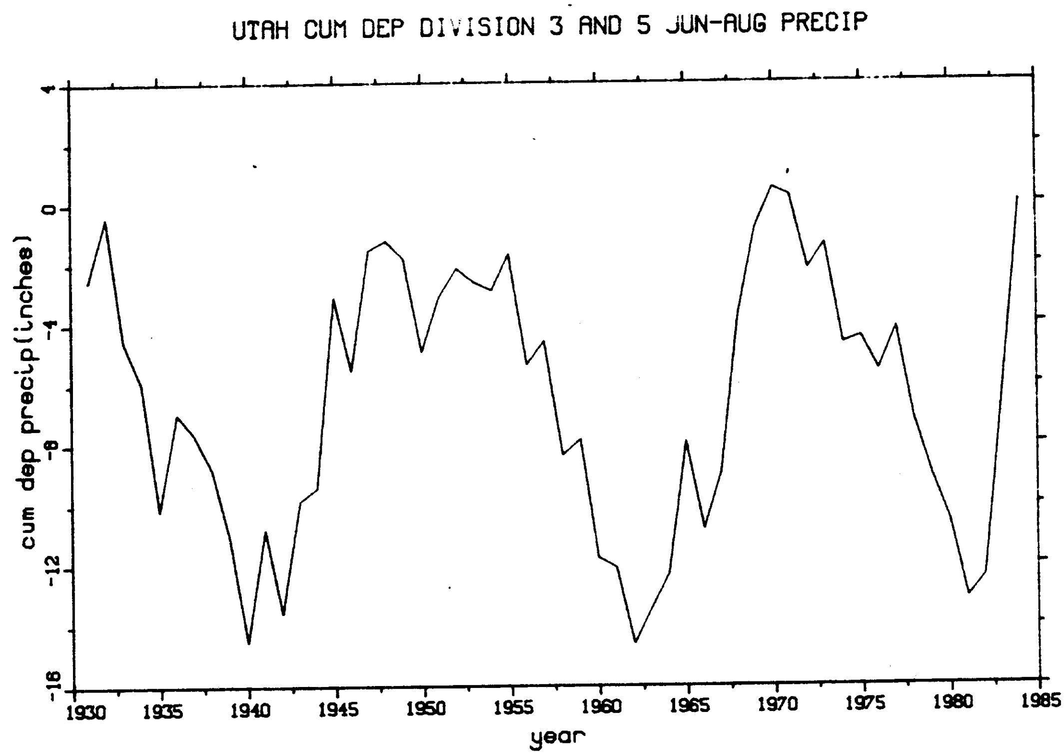
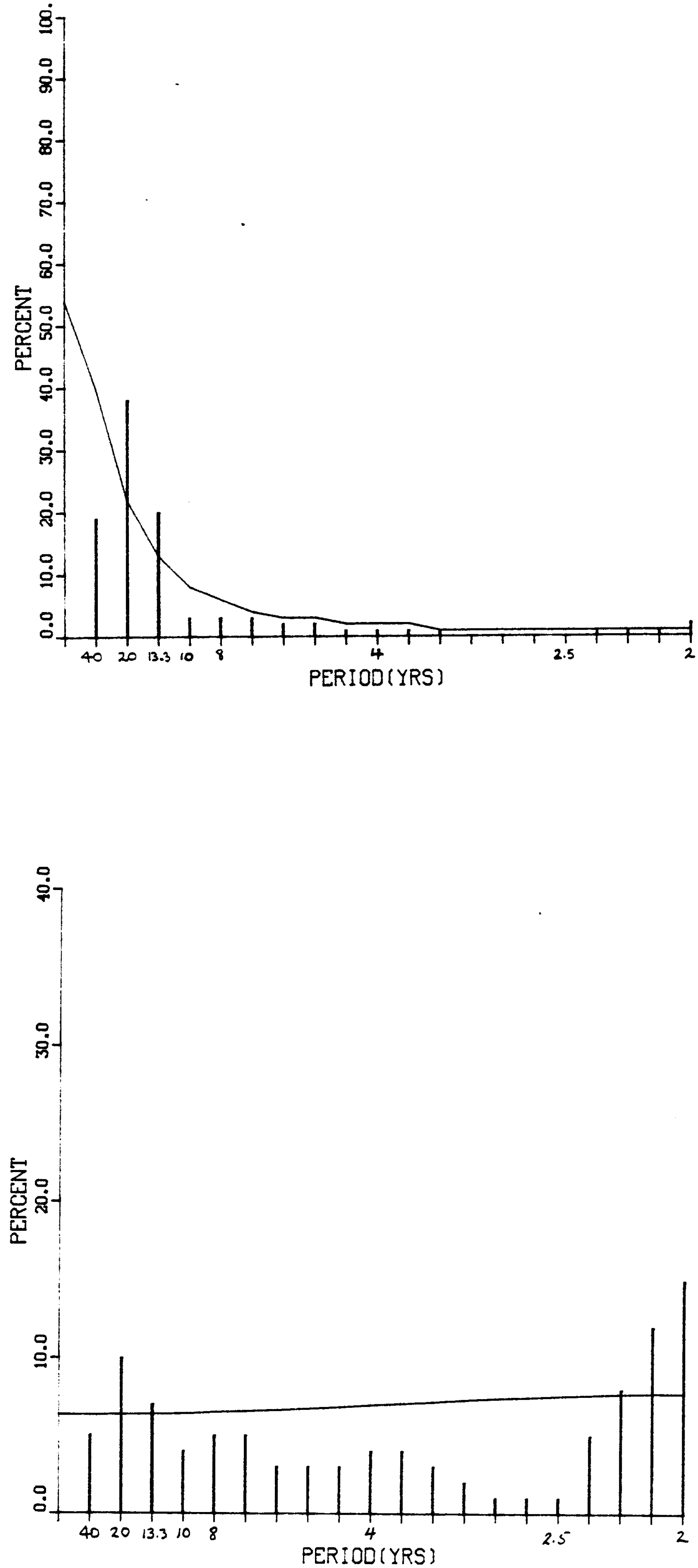


Figure 9 Variance spectrum of combined divisional summer precipitation a) of the cumulative departure curve given in Fig. 8. b) of the unsmoothed annual summer anomaly values.



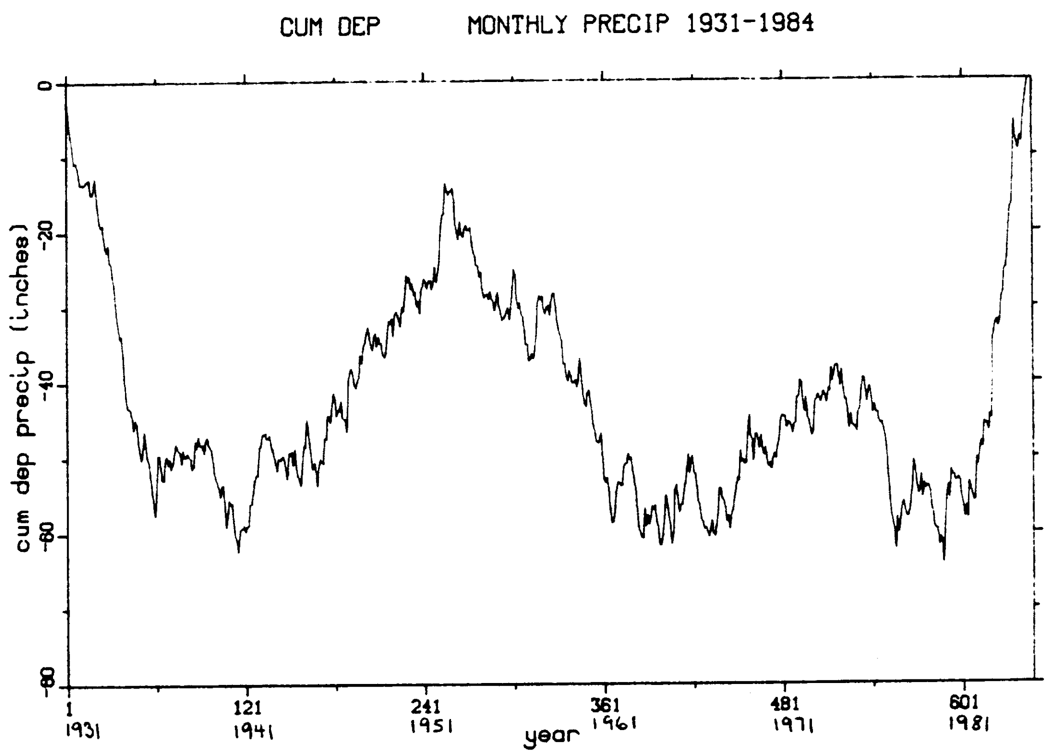


Figure 10 As in Fig. 4, except for all the monthly values from January 1931-September 1984 (645 months).

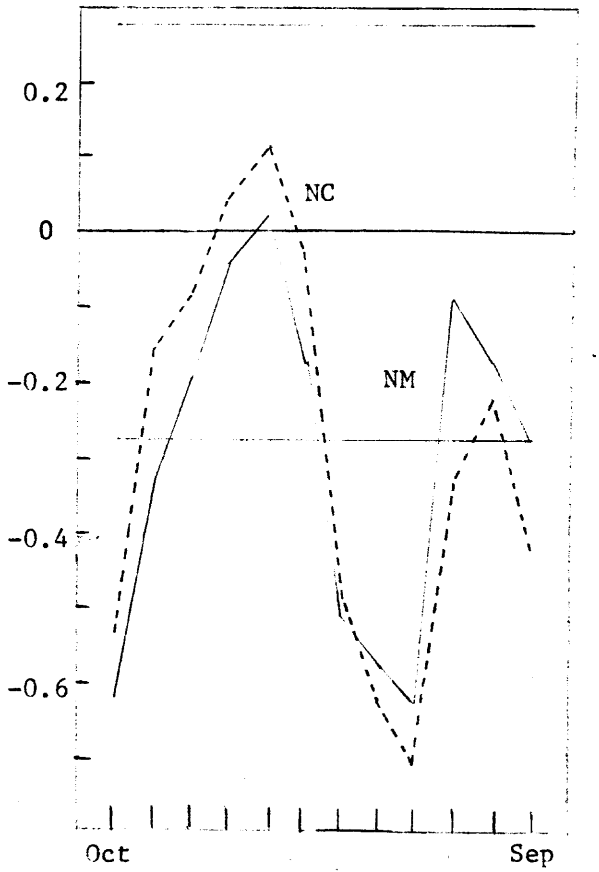


Fig. 11. Pearson correlation coefficients between temperature and precipitation for the two divisions.

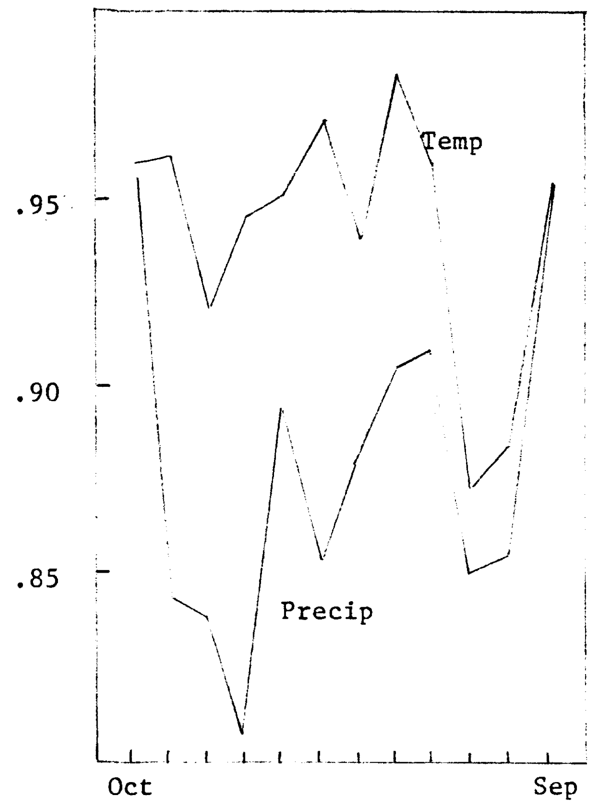
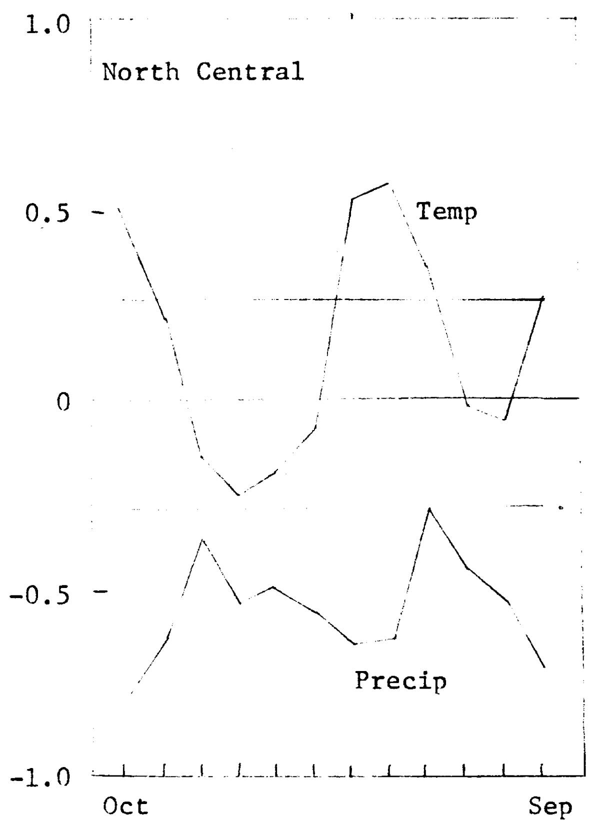


Fig. 12. Pearson correlations between temperature in the two divisions, and between precipitation in the two divisions.



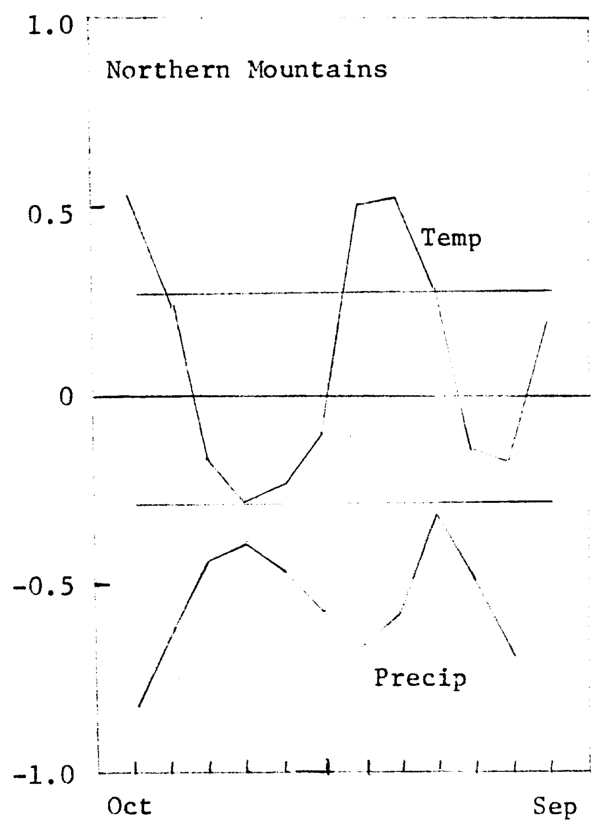


Fig. 13. Pearson correlations between percent possible sunshine and temperature and precipitation, for the two divisions.

TABLE 1. PEARSON CORRELATION COEFFICIENTS FOR SEASONAL PRECIPITATION AND TEMPERATURE

SEASONAL PRECIPITATION, NORTH CENTRAL AND NORTHERN MOUNTAINS DIVISIONS

SEASON	r	N	SIGNIFICANCE
Sep-May	0.81	54	0.01
Dec-Feb	0.81	53	0.01
Oct-Apr	0.78	53	0.01
Jun-Aug	0.91	54	0.01
Oct-Sep	0.84	53	0.01

SEASONAL TEMPERATURE AND PRECIPITATION

	NORTH CENTRAL DIVISION			NORTHERN MOUNTAINS DIVISION		
SEASON	r	N	SIGNIF.	r	N	SIGNIF.
Sep-May	-0.40	54	0.01	-0.35	54	0.05
Dec-Feb	-0.09	53	-	-0.22	53	-
Oct-Apr	-0.33	53	0.05	-0.35	53	0.05
Jun-Aug	-0.58	53	0.01	-0.30	53	0.05
Oct-Sep	-0.48	54	0.01	-0.39	53	0.01

TABLE 2. COMPARISON OF MEANS BETWEEN RECENT WET PERIOD AND PREVIOUS RECORD

DIVISION	t-STATISTIC*	DEGREES OF FREEDOM	2-TAIL PROBABILITY
North Central	3.76	24.86	0.001
Northern Mountains	4.12	25.13	0.004
Both Together	3.52	24.94	0.002

*Using separate variance estimates for populations with unequal variances (see Nie *et al.*, 1975, Statistical Package for the Social Sciences (SPSS), pp. 269-270).

TABLE 3. Regression results, showing values of standardized coefficients, coefficient of multiple determination, standard error of estimate, F-ratio for the regression model, and Mallows' C_p statistic.

DEPENDENT	INDEPENDENT	COEFFICIENTS	R^2	S.E.E.	F	C_p
		Standard Seasons, Divisions Combined				
Rise	T, MAM	-0.18				
	P, SON	0.60				
	P, DJF	0.43				
	P, MAM	0.20	0.73	0.43	31.24	4.25
Rise	T, MAM	-0.16				
	P, SON	0.59				
	P, DJF	0.45				
	P, MAM	0.19				
	Rise, lag 1	0.13	0.74	0.43	26.41	5.32
Fall	T, JJA	0.19				
	P, MAM	-0.29				
	P, JJA	-0.52	0.57	0.29	21.48	3.25
Fall	P, MAM	-0.19				
	P, JJA	-0.53				
	Rise	-0.40	0.68	0.25	33.82	2.42
		Accumulation and Loss Seasons, Divisions Separate				
Rise	P, Sep-May, NC	0.59				
	P, Jun-Aug, NC	0.26				
	P, Sep-May, NM	0.24				
	Sun, Jun-Aug	0.32	0.78	0.49	41.45	4.19
Fall	P, Jun-Aug, NC	-0.55				
	P, Sep-May, NM	-0.59	0.78	0.48	88.29	-5.09

SESSION 3

PHYSICAL FACTORS AND MODELS

An Introduction To Great Salt Lake Effect Snowfall

David M. Carpenter
National Weather Service Forecast Office
Salt Lake City, Utah

I. Introduction

A lake effect snowfall along the Wasatch Front is typically a post cold frontal event in which the duration of snowfall downwind of the lee shore is extended in cold, northwest flow. Where one would normally expect snowfall to end with the passage of the upper level trough, lake effect snows usually continue for several hours after that event.

This paper is based on data taken from lake effect cases that occurred in the northwest flow following the passage of an upper level trough. Occasionally it appears that a lake effect may have occurred in southwest flow, which enhances or produces snowfall close to the northeast shoreline. These cases are relatively infrequent and are not included in this study. The necessity of a rapid infusion of cold air, usually in the form of a cold pocket, and the fact that this air would have to come from a southerly direction makes a conventional lake effect hard to come by. What probably occurs in most cases when snowfall is heavier along the northeast shoreline is a warm advection pattern associated with a cut-off low over Nevada as described by Williams (1962). Terrain produced upslope and differential friction from the lake surface to the shoreline would enhance snowfall in that vicinity, but a contribution of instability and moisture from the Great Salt Lake to the air mass is probably not a factor in most of these cases. Even after limiting the study to storms that occur in northwest flow, the effect of the Great Salt Lake on snowfall is not one that can be easily measured.

Following are some ideas resulting from the study of the various elements which should contribute to Great Salt Lake effect snowfalls. Twenty one lake effect cases dating from March 1971 to October 1984 were used. The last case October 18, 1984, resulted in the heaviest 24 hour snowfall on record at the Salt Lake International airport. All of the cases produced four or more inches of snow at some point in the valley and seven of them produced over a foot. This set of data is not intended to represent a complete listing of all Great Salt Lake effect cases which occurred during this time period. Instead, it is more likely to be a representation of the most dramatic or most noticeable cases.

The percentage of annual snowfall in the Salt Lake Valley which may be attributable to lake effect is not known. However, the average number of lake effect cases per year has been estimated to be between 6 and 8 (Alder,1977). The average number of days per year in which 0.1 inch of snow or more is measured at the Salt Lake International Airport is 34 (Figgins,1984). Other studies indicate that lake effect from the Great Salt Lake probably occurs a rather small percentage of the time (Dunn, 1983). Considering these facts and the restrictive nature of the required temperatures and wind flows necessary to produce a lake effect, it is felt by most of the forecasting staff at the Salt Lake International Airport that the total contribution of lake effect snowfall to the annual snowfall of the Salt Lake Valley[1] and nearby areas is probably rather small. Nevertheless, the lake effect can produce dramatic results, and is the object of much interest and speculation among those who watch the weather along the Wasatch Front.

II. Elements of a Great Salt Lake effect Snowstorm

A. Topographical Influences

The Lake is literally surrounded by mountain ranges which at one time formed islands and the shoreline (Fig. 1). The most prominent of these north-south mountain ranges is the Wasatch Range. The Great Salt Lake is oriented approximately along the 320 degree radial from the Salt Lake International Airport. The longest fetch, or over water trajectory, is approximately 75 miles long along that 320 degree radial. This fetch is realized in northwest flow for the Salt Lake Valley and in northerly flow for the Tooele Valley. Both valleys slope upward away from the Lake shore and form a rough half bowl shape.

In addition to upslope, the topography makes two other modifications to the surface wind that are worthy of note. The first of these is illustrated in Figure 2. An exaggeration of a simple balance of forces is used to explain frictional convergence along the lee shore of

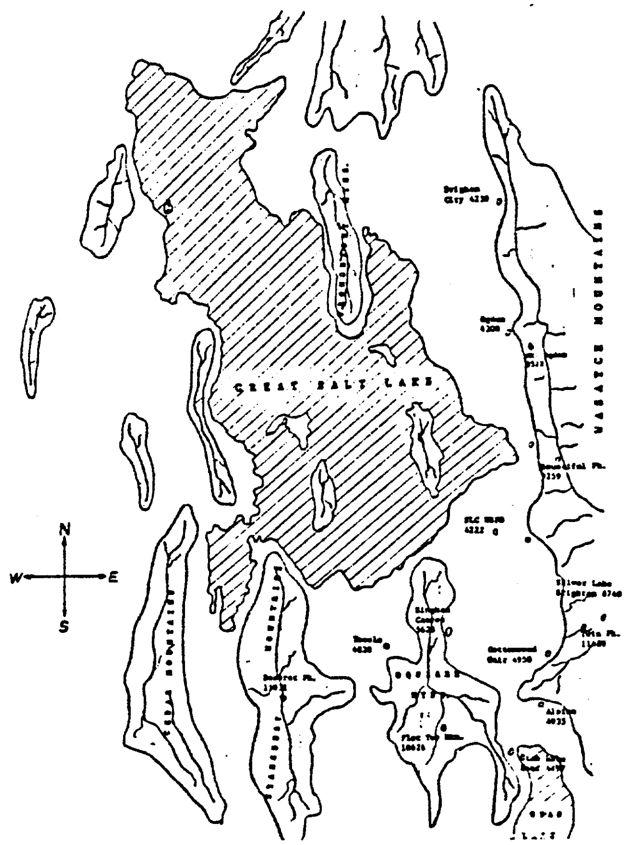


Figure 1 - Map of the Great Salt Lake and surrounding topographical features.

the Lake (Hess,1959). In reality the dynamics and kinematics of this process are much more complicated. At all four points, the pressure gradient force (PG) is the same, but the frictional drag force (F) varies with the roughness of the surface. Over the relatively smooth lake surface the frictional force is less than that over land. Therefore, the wind velocity (V) is greater over the lake than over land. This difference in speed is enough to cause an area of low level convergence along the lee shoreline (Lavoie,1972). The second modification to the low level flow is the channeling that occurs as air is pushed against the sides of the mountain ranges.

Figure 3 shows the resulting streamlines of a hypothetical case of

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[1] *The term Salt Lake Valley will be used to refer to the Salt Lake City metropolitan area including all municipalities near Salt Lake City and located west of the Wasatch Mountains and east of the Oquirrh Mountains, south of the Great Salt Lake and north of the Transverse Mountains. The Transverse Mountains are shown in Figure 1 as the point of the Mountain just north of Alpine.*

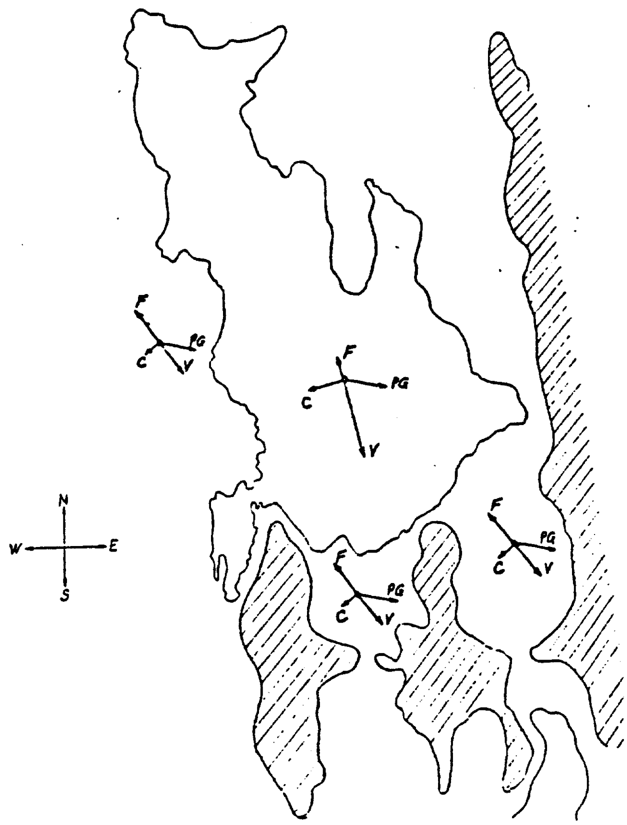

Figure 2 - A simple balance of forces illustration of the effect of differential friction on the surface wind flow in the vicinity of the Great Salt Lake.

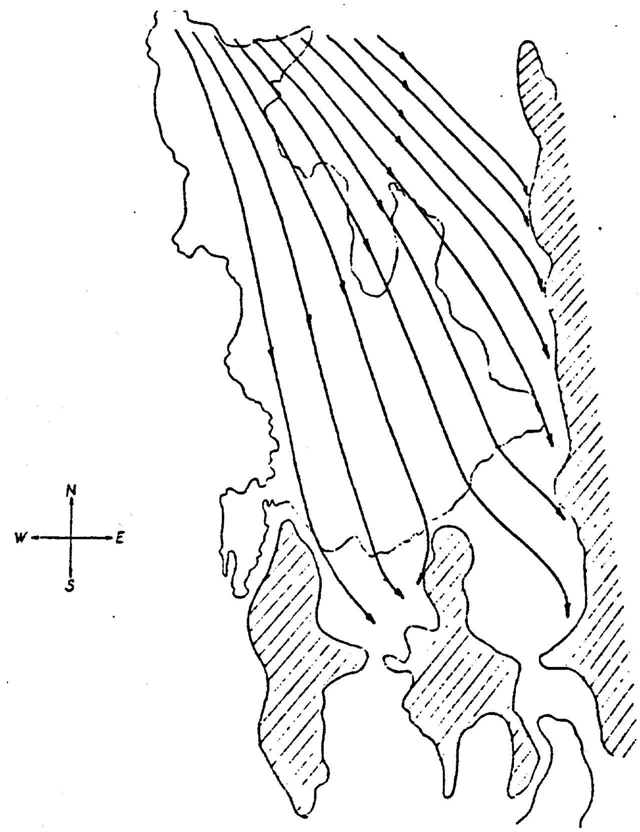

Figure 3 - An idealized illustration of the surface streamline analysis of northwest flow when the effects of differential friction and channelling by the Wasatch Front are taken into account.

northwest, low level flow. Note that the channeling effect and the backing of the wind as it moves onshore, combine to produce low level convergence in the upper portions of the Tooele Valley and along the east bench area of the Salt Lake Valley.

During some lake effect storms, the surface wind at the Salt Lake International Airport has been known to shift to the southwest and then to the southeast, while the wind at nearby surrounding stations remains northwesterly. This is possibly a reflection of something called a local front or boundary layer front by Garner (Garner,1983). This is something similar to a miniature New England Coastal Front (Bosart, Vaudo, and Helsdon, 1972).

In a Great Salt Lake local front, cold air at the surface becomes dammed up against the mountains in the upper portion of the valley and forms a small cold dome (Figure 4). Then, the low level flow coming off the Great Salt Lake rides up and over this cold dome, producing vertical motion. The colder air in the valley is maintained as being colder than the air coming off the Lake by convective down drafts and precipitation. Eventually with the help of a frictional disturbance along the lee shoreline, the flow in the valley becomes detached from the larger scale flow and can switch to a southerly direction. Without mesoscale data to confirm the above

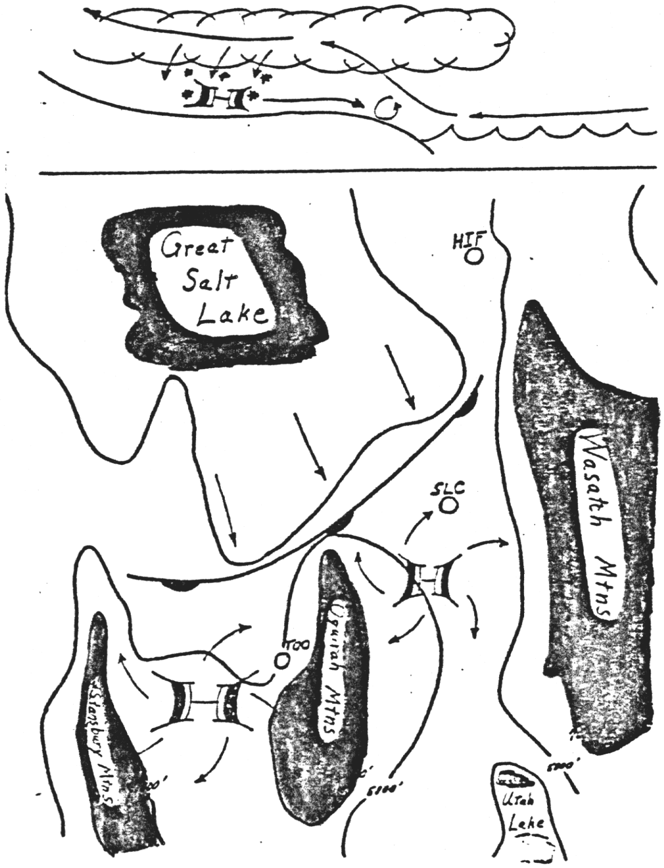

Figure 4 - Illustration of Great Salt Lake local front and cold air damming.

scenario, it is mostly conjecture, but not unreasonable.

B. Temperature of the Lake

The temperature of the Great Salt Lake is important to Lake Effect snowfall for two reasons: one, it indicates the amount of energy available in the surface of the Lake that may be transferred to the lowest layers of the atmosphere, to produce convection; and two, it determines the maximum allowable saturation vapor pressure of the air mass over the lake, for a transfer of moisture to occur between the Lake and the air mass (Ellenton and Danard, 1979). Since the Great Salt Lake brine has a reduced vapor pressure compared to fresh water, the second consideration cannot be overlooked. This will be dealt with in another section.

For the purpose of this study, it would be best to have a record of Great Salt Lake temperature measurements for several locations at the same time of the day over many years. For the purpose of real time forecasting, it would be best to have the same temperature data available in real time. However, neither of these situations is the case and we will have to settle for data taken at one point (the southern shore near the Salt Air Resort), twice a month by the United States Geological Survey. These data were then compared to the previous seven days actual mean air temperature at the Salt Lake International Airport (Figure 5) and to the normal mean air temperature of the airport (Figure 6). Two least squares curves were fitted to the data to predict the temperature of the Great Salt Lake. Figures 5 and 6 show the resulting least squares curves (solid lines with dots) and the distribution of data points about the curves (detached lines with dots). The correlation coefficient for the curve in figure 5 is 0.88 with a standard error of 4.78 degrees F. For figure 6, the correlation coefficient is 0.91 with a standard error of 4.21 degrees F.

For the 21 cases used in this study, the average Great Salt Lake temperature computed from the 7 day mean curve was 44.6. The average Lake temperature computed from the normal curve was 45.2. There is an obvious difference in the predicted Great Salt Lake temperature between the two curves. For example, using a temperature of 37 degrees F for the normal mean, produces a Lake temperature estimate of 42.6, while a 7 day mean temperature of 37 degrees F results in a Lake temperature estimate of 38.9 degrees F. Even though the difference is small in

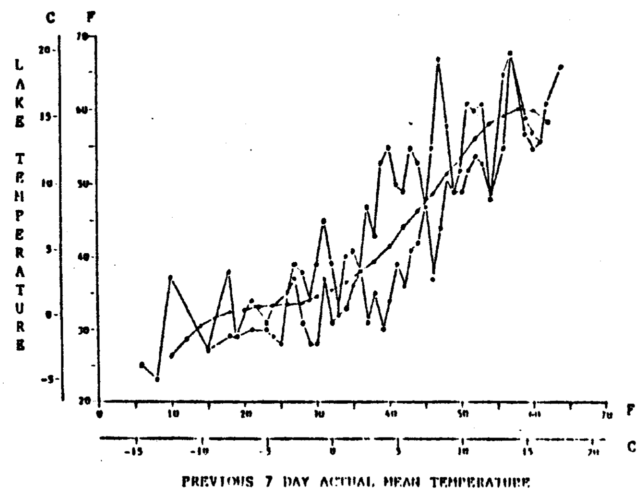

Figure 5 - Comparison of Great Salt Lake temperature verses the average of the previous 7 days actual mean air temperature for the Salt Lake City International Airport.

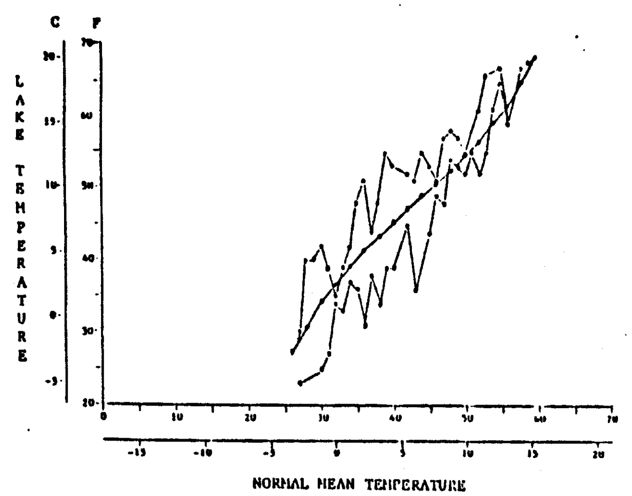

Figure 6 - Comparison of Great Salt Lake Temperature verses the normal mean temperature at the Salt Lake Airport for the date.

this example, it is easy to see that a period of abnormally warm or cold temperatures could cause a large difference in the two estimates.

Considering the size of the standard error for these two curves, as long as real time data on the temperature of the Lake is not available, an average of the two predictions will normally produce an acceptable prediction of the Great Salt Lake temperature. Averaging the two predictions will allow for an abnormally warm spell to be taken into account.

C. Instability Criteria

One of the necessary (but not sufficient) conditions for a significant Great Salt Lake effect snowfall is an unstable layer of air of sufficient depth to produce snow. This instability is achieved when the Lake is warm enough and/or the air mass is cold enough to create a lapse rate which is moist adiabatic or greater. A convenient way to measure this is to use the temperature difference between the Lake and 700 mb (Rothrock, 1969). For comparison, using a surface pressure of 870 mb (a typical surface pressure) and a Lake temperature of 5 degrees C, a moist adiabatic lapse rate results in a Lake to 700 mb temperature difference of around 10 degrees C and a dry adiabatic lapse rate in a difference of around 17 degrees C.

The optimum Lake to 700 mb temperature difference appeared to be between 17 and 23 degrees C. Differences much greater than this (25 degrees or more) seemed to cause a decrease in convective activity rather than an increase. Temperatures of minus 18 degrees C at 700 mb were not uncommon in these cases. This is probably due to several factors related to the cold nature of the air mass involved, but the most important of these is probably the fact that air which is that cold is generally very stable, and the lake effect convection is not strong enough to compensate for that stability.

D. Vapor Pressure Differential

The Great Salt Lake brine has a reduced vapor pressure compared to that of fresh water. A study was conducted to determine vapor pressure as a function of brine concentration and temperature (Dickson,Yepson, and Hales, 1965). The results are illustrated in figure 7. The first two curves, labeled $E_s(29.2\%)$ and $E_s(24.7\%)$, show the resulting vapor pressure vs. temperature curves for the Lake for two values of brine concentration. The curve labeled E_s is the same distribution for distilled water.

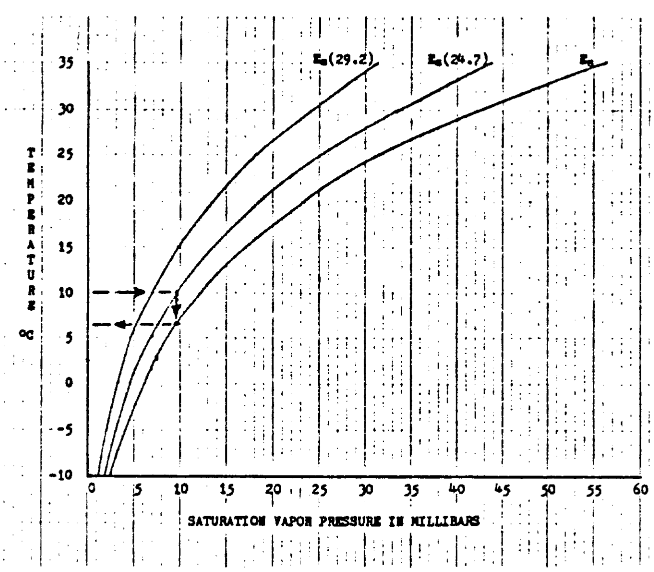

Figure 7 - Saturation vapor pressure taken over Great Salt Lake brine for varying degrees of salinity and over distilled water verses the temperature. Taken from Dickson, Yepson Hales, 1965).

A necessary requirement for a transfer of moisture from the Lake to the air mass, is that the vapor pressure over the Great Salt Lake brine must be greater than the vapor pressure of the overlying air mass. For example using the point on the 24.7% curve ($E_s(24.7)$) corresponding to a Lake temperature of 10 degrees C, we follow a constant vapor pressure line down to the distilled water curve (E_s) and then read the temperature on the scale on the left. The dewpoint temperature of the overlying air mass must be less than this temperature for a transfer of moisture to occur between the Lake and the air mass. In this case the dewpoint temperature of the air mass must be less than about 6.5 degrees C. Given the condition that the salinity of the Great Salt Lake is probably less than even 24.7% , a rough rule of thumb could be that a Lake to dewpoint temperature difference of 5 degrees C or greater would satisfy this condition.

The change in the precipitable water value in the downstream air mass as a result of Lake Effect is probably only a few hundredths of an inch. However since we are dealing with precipitable water values in the .10 to .40 inch range, and considering that the vertical motions produced by the terrain effects and convection can be considerable, the effect of this increase is worth mentioning (Harley, 1965).

If one were to assign all the elements of Great Salt Lake lake effect snow, a degree of importance in contributing to the rate of snowfall, it is felt that the vertical motion produced by low level convergence and upslope and the addition of sensible heat to the lower levels of the atmosphere, should receive a higher ranking than the addition of moisture by the lake.

E. Inversion Height

A limiting factor to Lake Effect snowfall and one which is strongly correlated to the end of the snowfall is the height of the subsidence inversion following the 700 mb trough (Rothrock, 1969). The height of this inversion serves to limit the vertical extent of the convective cloud tops. Rawinsonde traces from the 21 cases in this study indicate that an inversion height between 650 and 700 mb is probably the minimum for production of any precipitation. Once the inversion drops below 700 mb most storms are more or less over.

F. Weak Upper Disturbances

One ingredient which is apparently necessary for a significant Great Salt Lake effect snowfall is the presence of a weak upper level disturbance. This disturbance is often not shown in the operational vorticity analyses due to the large spacing between upper air stations. Most of the time a temperature trough will be the only indication of its presence in the upper level charts. The surface pressure tendency may also reflect its presence (Dunn 1983).

G. Location of Snowfall vs. Wind Direction

The data confirms the assumption that the direction of wind flow determines the location of the heaviest concentration of snow. The areas which received a discernable concentration of heavy snow during a lake effect were for the most part limited to four locations. These are shown in Figure 8. Area 1 appears to receive the most Lake Effect snowfall. Area 2 is not well covered with climatological stations, and therefore will not be represented very

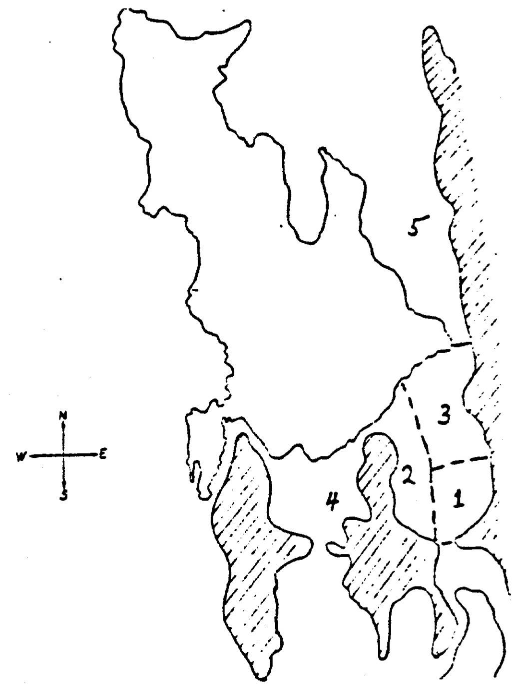

Figure 8 - Areas which receive lake effect snow.

well in the following discussion of that data. However notes taken at the Salt Lake City forecast office show a strong tendency for lake effect in this area. The third area includes the east foothill area of Salt Lake City and the valley floor from about 2100 south northward. This is possibly just an extension of the first area northward, but there does appear to be a separate response to wind direction. The fourth area is in the Tooele Valley near the town of Tooele. The fifth area is from Davis county northward, but as was mentioned before, no cases for this situation are included in this study.

Figure 9 is included to allow a comparison to the normal water equivalent precipitation for the months October through April. The remaining maps (figures 10 through 17) are cumulative water equivalent, storm total precipitations for groupings of lake effect storms. The intention is to show a pattern of snowfall rather than to imply a quantitative average of some kind.

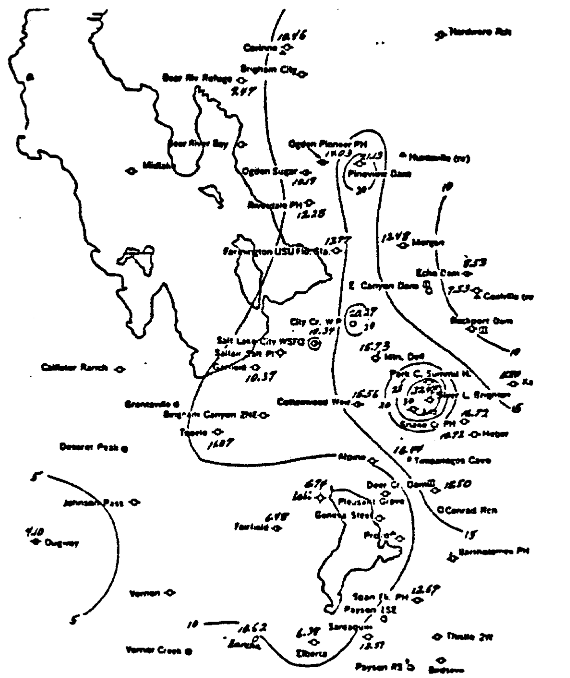

Figure 9 - October through April normals for water equivalent precipitation in the vicinity of the Great Salt Lake.

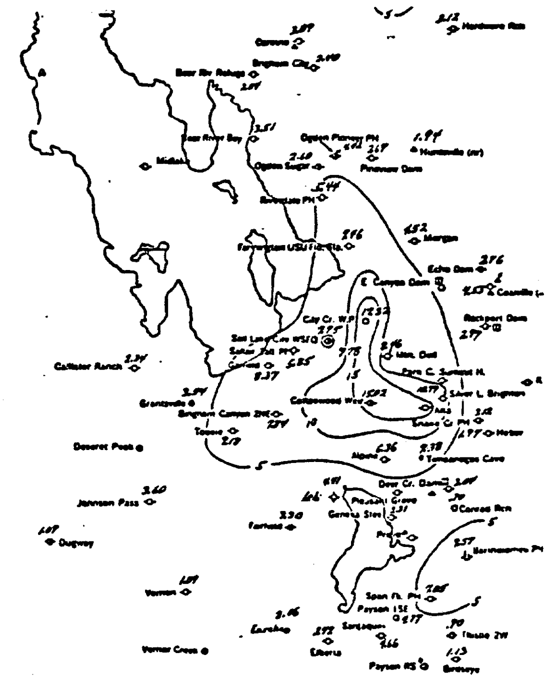

Figure 10 - Total water equivalent precipitation for all of the 21 Lake Effect snowfall cases in the study.

The 700 mb wind direction was used for two reasons; one, it is convenient to obtain since it is routinely observed and, it is not as subject to frequent fluctuations due to terrain or local drainage effects.

Data are only presented for situations in which three or more cases would apply. In figures 11 through 17, each case was categorized using the average 700 mb wind direction, during the Lake Effect. Many cases are included in two maps because their average 700 mb wind direction was an even multiple of ten and thus fell on the line between two categories. Figure 10 shows the cumulative precipitation for all Lake Effect cases. Note that the area of maximum snowfall extends approximately from Bountiful southward to Sandy, and has been drawn to include Little Cottonwood Canyon to Alta. This comprises areas 1 and 3 mentioned above. A complete set of data for Alta was not available in the monthly Utah Climatological Data Publication, which was used in this study. However, it is felt by the author that Alta would easily fall within the 15 inch isopleth.

Figures 11 and 12 show the precipitation data for cases with an average 700 mb wind direction from 330 to 350 degrees inclusive. The bull's eye of heavy snow is located in the Tooele Valley, or area 4 as described above. Area 2, or the western portion of the Salt Lake Valley also receives heavy snowfall in many of these cases. A dramatic switch in the location of the bull's eye occurs when the 700 mb wind direction becomes more westerly. This is shown in Figure 13 with the average 700 mb wind direction from 320 to 330 degrees (inclusive). The heavy snowfall bull's eye in this case is located in the southern portion of the Salt Lake Valley, or area 1. Average 700 mb wind directions from 290 to 320 degrees produce essentially the same results (figures

13 through 16). As the wind direction becomes even more westerly as in figure 17, a new maximum center appears to develop along the east bench area of the city or area three. Additional information on the frequency and amount of precipitation for certain stations in the area, with respect to 700 mb flow direction and instability, was developed by Elliott, Thompson, and Griffith (1985).

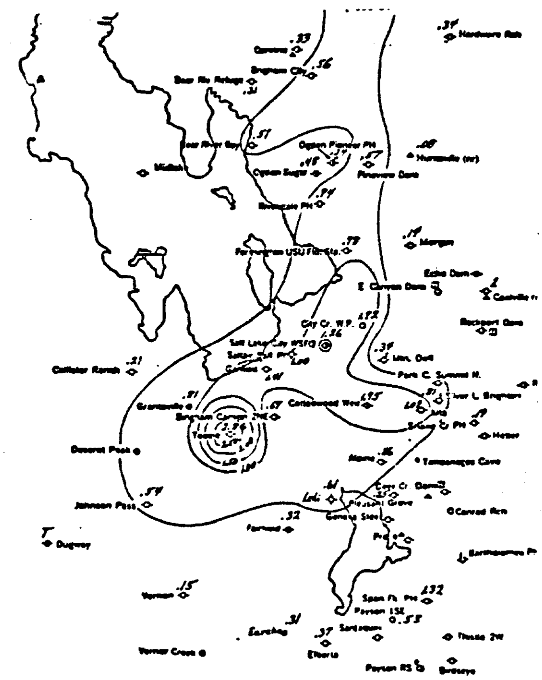

Figure 12 - Total water equivalent precipitation for all Lake cases in which the average 700 mb wind direction was from 330 to 340 degrees (inclusive).

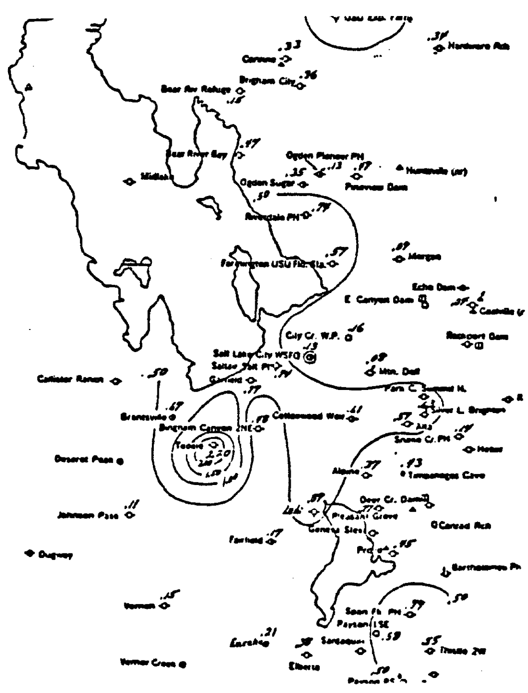

Figure 11 - Total water equivalent precipitation for all Lake Effect cases in which the average 700 mb wind direction was from 340 to 350 degrees (inclusive).

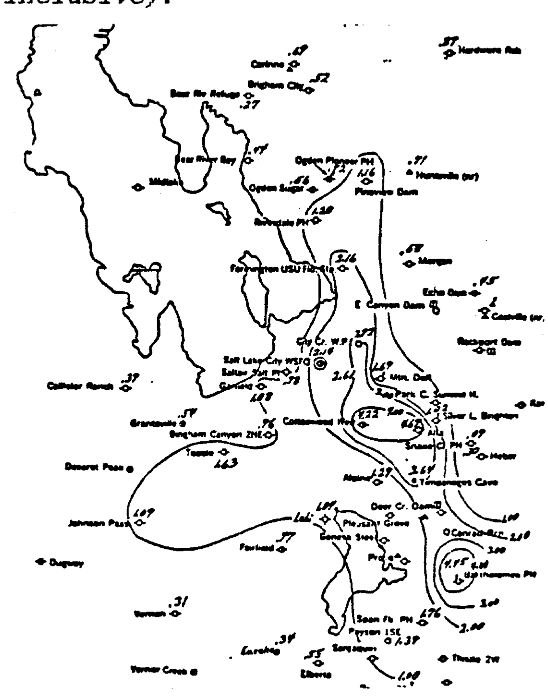

Figure 13 - Total water equivalent precipitation for all Lake cases in which the average 700 mb wind direction was from 320 to 330 degrees (inclusive).

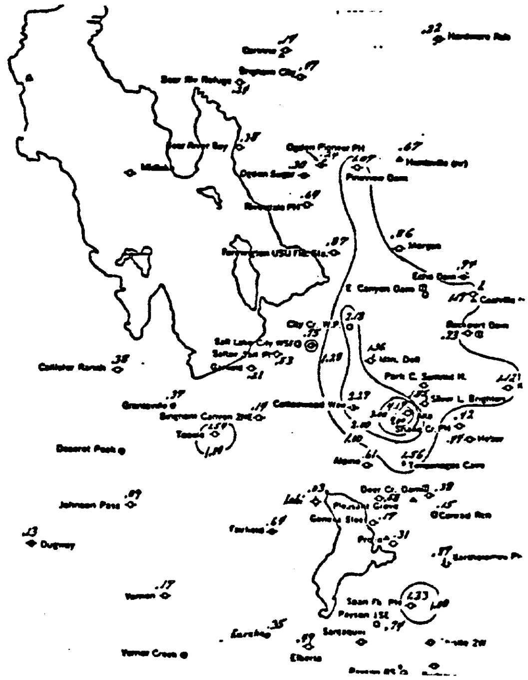

Figure 14 - Total water equivalent precipitation for all Lake cases in which the average 700 mb wind direction was from 310 to 320 degrees (inclusive).

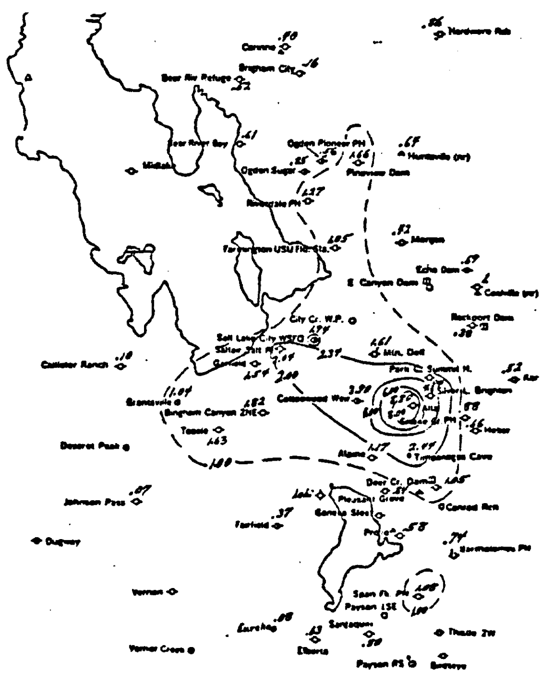

Figure 15 - Total water equivalent precipitation for all Lake cases in which the average 700 mb wind direction was from 300 to 310 degrees (inclusive).

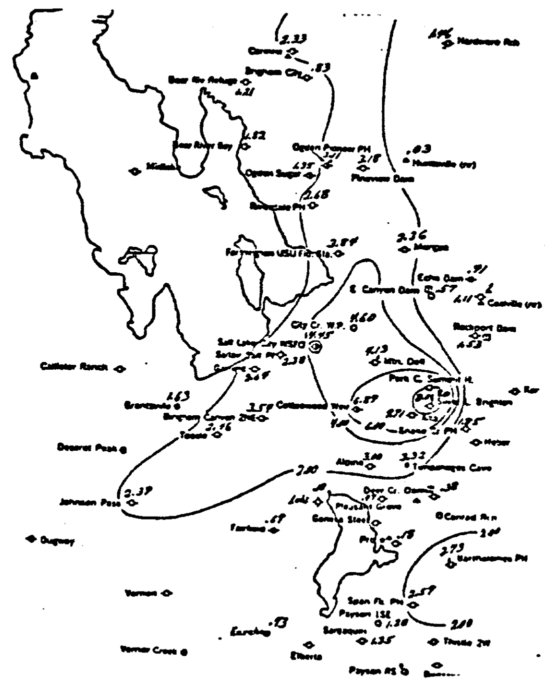

Figure 16 - Total water equivalent precipitation for all Lake cases in which the average 700 mb wind direction was from 290 to 300 degrees (inclusive).

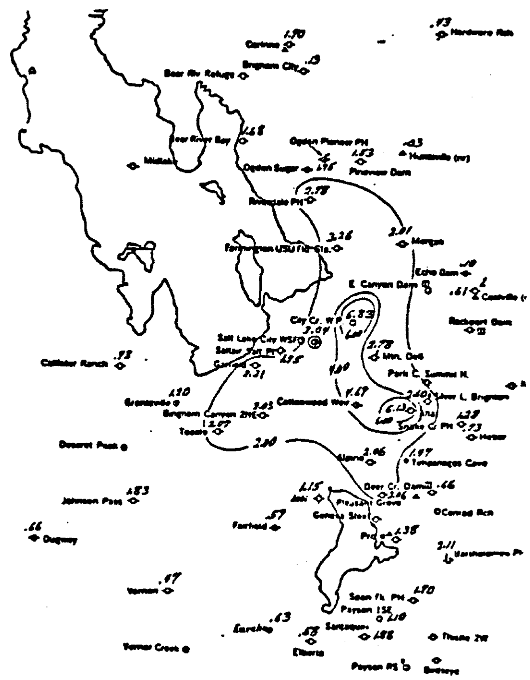

Figure 17 - Total water equivalent precipitation for all Lake cases in which the average 700 mb wind direction was from 280 to 290 degrees (inclusive).

IV. Summary

The conditions which are apparently necessary for Great Salt Lake lake effect snowfall could be summarized as follows. The average 700 mb wind direction should be between 270 and 360 degrees for most cases. The optimum range for the difference between the temperature of the Great Salt Lake and the average 700 mb temperature is between 17 and 23 degrees Celsius. The difference between the temperature of the Great Salt Lake and the upstream dewpoint temperature should be at least 5 degrees C. Some kind of upper level support such as a temperature trough or a jet streak or both should be available for a significant snowfall.

V. Acknowledgments

This study is the brain child of William Alder, who saw the need for more information on Lake Effect snowfall and provided much advise and technical guidance. Thanks also to those who reviewed earlier drafts and made comments and corrections, including Western Region Scientific Services Division and the staff of the forecast office. I am grateful to Greg Smith who spent many hours transcribing data for the cases. Thanks to Dr. Joseph Schaefer for the loan of several years of data on microfilm.

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Models for Evaluating the Effects of a Causeway and Dikes on the Water and Salt Balance of Great Salt Lake, Utah

by K. M. Waddell

U.S. Geological Survey
Salt Lake City, Utah

During 1957-59, a permeable rock-fill causeway was constructed across Great Salt Lake by the Southern Pacific Transportation Co. Although the causeway contained two 15-foot wide culverts, it functioned as a partial dike dividing the lake into north and south parts. The causeway interrupted the formerly free movement of brine within the lake. This, along with the fact that more than 95 percent of the freshwater surface inflow enters the lake south of the causeway, resulted in substantial changes in the hydrology and chemistry of the lake. The salinity of the south part was diluted, and the north part became so concentrated that large quantities of sodium chloride were being deposited in that part.

During the 1970s, the U.S. Geological Survey, in cooperation with the Utah Geological and Mineral Survey and Utah Division of Water Resources, developed two models for predicting the water and salt balance of Great Salt Lake. The first model was used to evaluate the effects of the Southern Pacific Transportation Co. causeway on the lake and determine the width of additional culverts needed to achieve different degrees of salt equilibrium between the north and south parts. It was determined that a culvert width of 500±100 feet would be required to cause the salt concentration in the south part to become within about 85 percent of that in the north part. Also, widening of culverts in excess of 500 feet would result in little additional gain of salt concentration in the south part.

The second model was developed for predicting the water and salt balance for various diking options in the Great Salt Lake. The model provides for nine diking options. These include combinations of eight areas east of a line joining Antelope Island, Fremont Island, and the Promontory Mountains. Another option includes the part of Great Salt Lake that lies north of the Southern Pacific Transportation Co. causeway.

The model treats the salt balance of the diked areas from the standpoint of inflow-outflow balance with complete mixing. Also, no allowances are made for any stratification or chemical changes due to interaction with the sediments, or for solution of entrapped brines or residual salts. Because the degree of inaccurracy created by these assumptions is not known, the concentrations predicted by the model are not regarded as absolute, but as relative indexes by which to compare the various diking alternatives.

The equations and physical relationships used in the existing models need to be verified and or modified for current (1985) lake levels. Most of the data used for development of the models was collected when the lake was below an altitude of about 4200 feet or about 9 feet below the 1984 peak. During 1984, considerable detailed data were collected to evaluate the effects of a 270-foot wide causeway breach. These and other data collected during the previous years could be used to update the models to accommodate the current (higher) lake levels.

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GREAT SALT LAKE WATER BALANCE MODEL

Norman E. Stauffer, Jr

Chief of Hydrology & Computer Applications
Utah Department of Natural Resources/Water Resources

INTRODUCTION AND HISTORY

State and Federal Water Resources planning agencies have for several years prepared mean annual water budgets of the Great Salt Lake for water planning purposes. The first results of using a computer model for a water budget analysis of the Great Salt Lake was were published by Steed and Glenne (1972) at the University of Utah. This model was based on mean monthly historic flows for the 1944-1970 time period. Precipitation and evaporation estimates were obtained from climatological stations near the lake. The result of this model was a refinement of a water budget for the Great Salt Lake based on mean monthly values for a 26 year period.

To facilitate the analysis of the hydrologic system of the Great Salt Lake over a much longer time period with a minimal amount of input data, the Utah Division of Water Resources (1974) developed a water balance model of the Great Salt Lake based only on annual input data. The model was able to predict June 1st as well as end of water year lake elevations for present water use conditions and with projected future upstream depletions to the system for the 1901-1973 hydrologic period. This model was later modified by the Division (1977) to simulate removal of water by pumping from the lake during high lake stages and the data base was expanded to include the 1851 through 1975 hydrologic period. The basic data required by the program included the annual inflow of the Bear, Weber, and Jordan Rivers, precipitation at the Salt Lake Airport, and elevation-area-volume data for the lake. Ungaged inflow and groundwater inflow were computed as a function of gaged inflow with the latter including lag-times up to three years. The Salt Lake Airport annual precipitation data were adjusted to estimate mean lake precipitation. Evaporation was computed assuming a constant evaporation rate of 4.33 feet per year for fresh water and adjusted as a function of salinity.

In 1971 the U. S. Geological Survey, in cooperation with the Utah Division of Water Resources, began a seven year study to monitor the parameters controlling the water and salt balance of the Great Salt Lake. The parameters included ground and surface water inflow, precipitation on the water surface, outflow from evaporation, and chemical quality of the lake and surface water inflow. In 1974 Waddell and Fields (1977) of the U. S. Geological Survey developed a computer model on the basis of preliminary inflow data. The model was updated by Waddell and Barton (1980) using additional data collected for calibration of the model. The model simulates both the hydraulic and salinity effects of the Southern Pacific Transporation Company Railroad Causeway and computes the lake levels and lake salinities for the 1931-1976 period on a monthly basis. The model can also be used to evaluate the water and salt balance for various combinations of diked bay areas of Great Salt

Lake such as Farmington Bay. The monthly input data for the model is available for only the 1931-1976 period in which the lake stage was at a relatively low level. It would be almost impossible to extend the data base backward in time to the 1870's when the lake was at its highest level. However, since the level of the Great Salt Lake is now approaching the historic high, it may be desirable and would be possible to extend the input data forward in time to the present allowing the model to be used to evaluate lake level control alternatives, diking alternatives, and effects of water development projects in the Great Salt Lake Drainage Basin.

Also during the 1970's, James et al. (1979) at the Utah Water Research Laboratory developed a stochastic hydrologic simulation model for the Great Salt Lake. They used the basic water balance equation developed by the Utah Division of Water Resources. The surface inflows, precipitation, and evaporation rate were all generated using techniques of operational hydrology employing a multivariate stochastic model. The stochastic hydrologic simulation model was coupled with a damage simulation model. The combined model provided the capability to estimate both water surface elevation probabilities and associated damages. The model was also used to analyze lake control alternatives.

With the recent rising levels of the Great Salt Lake, the Utah Division of Water Resources and the Utah Water Research Laboratory in 1983 jointly updated the water balance model, stage damage model, and stochastic generation model. The water balance model was modified to include the effects of the Southern Pacific Railroad Causeway and recalibrated using data for the 1944-1983 period. The "new" model was used extensively for studying lake control alternatives. A summary of the results of this effort was published by James et. al. (1984) at the Utah Water Research Laboratory.

WATER BALANCE MODEL

The lake annual water balance model developed by the Utah Division of Water Resources for the Great Salt Lake is described below. The form of the model is generally applicable to terminal lakes. The basic relationships for the model are the water balance equations:

$$V_t = V_{t-1} + QS_t + S_t + G_t + (P_t - E_t)AA_t \qquad (1)$$

$$VP_t = V_{t-1} + C1(QS_t + S_t + G_t) + (C3P_t - C2E_t)A_{t-1} \qquad (2)$$

where

V_t = volume of lake at the end of the t^{th} water year in acre-feet

VP_t = volume of lake at the peak annual stage in the t^{th} water year in acre-feet

in which

V_{t-1} = volume of lake at the end of the $(t-1)^{th}$ water year in acre-feet

QS_t = total gaged surface inflow of the Bear, Weber, and Jordan rivers and the outflow from Willard Bay to the lake in the t^{th} water year in acre-feet

S_t = ungaged surface inflow from tributaries around the lake in the t^{th} water year in acre-feet

G_t = subsurface inflow during the t^{th} water year in acre-feet

P_t = precipitation on the lake during the t^{th} water year in feet

E_t = evaporation rate from the lake during the t^{th} water year in feet

AA_t = average of the lake surface areas at the end of the $(t-1)^{th}$ water year and at the peak state of the t^{th} water year in acres

A_{t-1} = lake surface area at the end of the $(t-1)^{th}$ water year in acres

C1 = fraction of surface and subsurface inflow which occur before the peak stage

C2 = fraction of annual evaporation which occurs before the peak stage

C3 = fraction of the annual precipitation which occurs before the peak stage

In applying Equations 1 and 2, the initial stage can be translated into corresponding values for the lake surface area A_{t-1} and volume V_{t-1} which are the previous end of water year lake surface area and volume using lake stage-area-volume relationships. Annual values for the inflows, precipitation, and evaporation are first used in Equation 2 to compute the annual peak volume VP_t from which the peak lake surface area can be determined and averaged to obtain AA_t. Equation 1 is then used to compute the end of year lake volume V_t. For the next year V_t becomes V_{t-1} and the process can proceed iteratively for as many years as inflow, precipitation, and evaporation data are provided.

In the early versions of the model, the coefficients C1, C2, and C3 which are the fractions of inflow, evaporation and precipitation occurring before the peak lake stage had values of .75, .30, and .71 respectively. These values were based on the assumption that the lake would always peak on the first of June each year. The time of the lake peak is dependent on the inflow, evaporation, precipitation, and lake stage. These fractions are now computed for each year using this information as follows:

$$C1 = .632 + .0094\ QS_t/A_{t-1} + .0240\ EP_t \qquad (3)$$

$$C2 = .574 + .0309\ QS_t/A_{t-1} - .0836\ EP_t + .182\ P_t \qquad (4)$$

$$C3 = .972 + .0923\ QS_t/A_{t-1} - .0643\ EP_t - .110\ P_t,\ C3 \leq .97 \qquad (5)$$

where EP_t is pan evaporation in the t^{th} water year. The coefficients in equations 3, 4, and 5 were computed using a multiple regression analysis based on historic data for the 1944-1983 period.

The ungaged surface (S_t) and subsurface (G_t) inflows are estimated by the following relationships:

$$S_t = (0.296\ P_t - 0.0226\ E_t)AA_t \tag{6}$$

$$G_t = 0.015\ QS_t + 0.015\ QS_{t-1} + 0.015\ QS_{t-2} \tag{7}$$

The evaporation is input to the model as pan evaporation (EP_t) at the Bear River Refuge. Freshwater equivalent lake evaporation (ET_t) is computed as follows:

$$ET_t = 0.79655\ EP_t \tag{8}$$

The reduction in evaporation caused by salinity is estimated as outlined by James et al. (1979) with the following equations:

$$E_t = ET_t(1.0 - 0.00833C_t) \tag{9}$$

$$C_t = 0.12345\ WS/V_t,\ C_t \leq 27.5 \tag{10}$$

in which C_t is the mean lake salinity in percent by weight and WS is the total weight of salt in the lake in tons. The coefficients in Equation 7 were determined from data in Waddell and Barton (1980). The coefficients in Equations 6 and 8 were determined in calibration of the model for the 1944-1983 period using MINPACK-1, an optimization package developed by the Argonne National Laboratory by More et al. (1980).

To model the effects of the railroad causeway on the lake levels, a multiple regression analysis was made which resulted in the following equations:

$$DHP = -\ 547.43 + .13702\ ELV_{t-1} + .45190\ QS_t \tag{11}$$

$$DHL = -\ 320.20 + .07636\ ELV_{t-1} + .39097\ QS_t \tag{12}$$

where DHP and DHL are respectively the annual peak and end of water year head differences across the causeway. DHP is limited between 0.5 and 4.0 feet and DHL is limited between 0.3 and 3.5 feet. ELV_{t-1} is the $t\text{-}1^{th}$ end of water year elevation. To adjust the South Arm elevations for the head differences, DHP and DHL are multiplied by 0.375 and added to the South Arm annual peak elevation (ELP_t) and end of water year elevation (ELV_t). With the causeway now breached it is estimated the values DHP and DHL will be about 40 percent of the values computed by Equations 11 and 12.

MODEL USE AND RESULTS

The required input data for the model includes gaged river inflow, lake precipitation, pan evaporation and elevation-area-volume data. The elevation-area-volume data is shown on Figure 1. The gaged or correlated river inflow, lake precipitation and pan evaporation are shown in Table 1 for the 1853-1984 period. Inflows for both the historical and present modified inflows (adjusted over history with present level of water use) are shown in Table 1 as well as the historical annual peak lake level. For a detailed description of the data in Table 1 see James et al. (1985). Results from the model compared to the historical elevations

FIGURE 1. Great Salt Lake Elevation—Area—Volume Curves

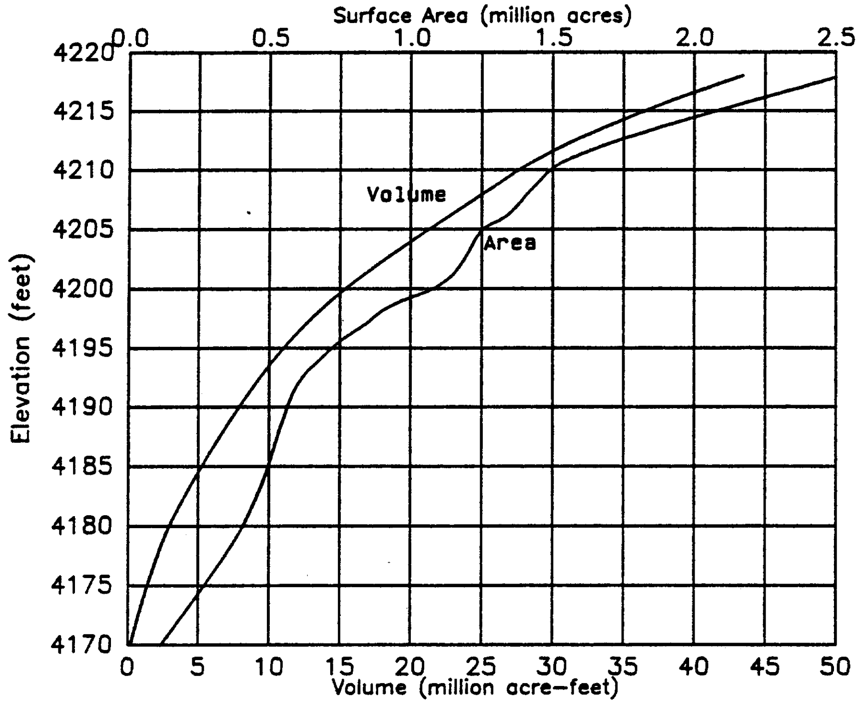


FIGURE 2. Great Salt Lake Water Balance Model Calibration Error

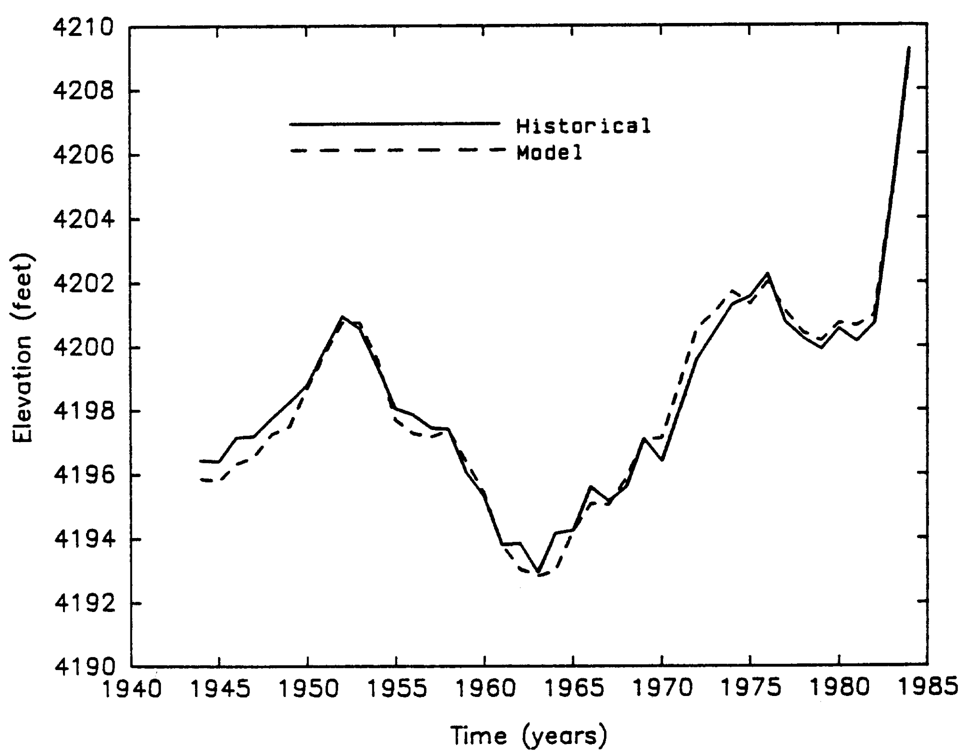


Table 1. Water Balance Model Input Data and Historical Annual Peak Elevation.

Water Year	Historical River Inflow (acre-feet)	Modified River Inflow (acre-feet)	Lake Precipitation (inches)	Pan Evaporation (inches)	Annual Peak Lake Level (feet-msl)
1851	2781800	1402900	8.27	64.42	4202.10
1852	4926200	3474200	6.56	68.26	4203.00
1853	3511400	2083400	12.60	57.94	4204.00
1854	3673000	2269000	9.80	61.73	4204.20
1855	2582400	1351200	8.67	63.66	4204.60
1856	2556100	1349100	12.13	58.50	4204.20
1857	627100	396000	14.66	55.75	4204.00
1858	1555100	870500	10.60	60.52	4203.00
1859	1001100	604500	14.10	56.31	4201.80
1860	2084800	1157400	9.50	62.21	4201.30
1861	2372000	1311000	10.51	60.65	4201.00
1862	4417200	3187200	12.54	58.01	4203.00
1863	4661800	3452800	7.33	66.39	4204.00
1864	3397400	2209400	10.28	60.99	4204.50
1865	4189200	3021200	13.28	57.17	4205.40
1866	3054500	1906500	19.30	51.98	4207.00
1867	4178800	3050800	13.68	56.74	4208.40
1868	4092200	2984200	16.24	54.32	4210.00
1869	4022200	2935200	11.50	59.29	4211.00
1870	2052400	1242700	13.84	56.57	4210.60
1871	4938900	3890900	8.18	64.59	4210.50
1872	5546000	4536000	9.85	61.65	4211.40
1873	2870900	1861900	13.01	57.47	4211.60
1874	2936800	1947800	9.28	62.58	4211.30
1875	2025300	1277600	12.27	58.78	4211.00
1876	3857000	2907000	15.38	54.58	4210.90
1877	3111300	2181300	8.55	63.79	4210.40
1878	2071600	1330800	11.35	60.26	4209.40
1879	2106300	1358200	5.99	68.59	4208.10
1880	2997800	2127800	8.16	61.45	4206.50
1881	3380000	2528000	10.50	62.22	4206.50
1882	2764200	1930200	8.49	62.07	4206.00
1883	3186200	1686200	7.53	65.25	4205.00
1884	3455000	2658000	15.35	52.52	4205.90
1885	3655600	2876600	13.76	56.54	4207.30
1886	3477100	2717100	11.61	61.27	4207.70
1887	2919600	2178600	8.21	64.97	4207.20
1888	2484700	1761700	7.16	67.20	4206.10
1889	994900	895400	7.83	66.97	4204.80
1890	2540300	1855300	14.79	55.32	4204.10
1891	1889400	1360700	12.55	56.43	4203.50
1892	1126200	987600	10.48	60.63	4202.90
1893	2274700	1640700	11.51	57.33	4203.00
1894	3761500	3144500	9.84	60.84	4203.00
1895	2019300	1459700	7.05	64.62	4202.20
1896	2841700	2257700	8.78	62.05	4201.80
1897	2685800	2116800	9.92	61.63	4202.30
1898	2394100	1841200	9.12	62.65	4201.90
1899	2566600	2028600	8.74	60.38	4201.20
1900	2065300	1543300	7.18	69.15	4201.00
1901	2022300	1507300	10.02	65.16	4200.00
1902	1226400	1109700	7.73	65.53	4199.00
1903	1510900	1255500	8.37	63.56	4197.70
1904	2632500	2140500	10.03	62.58	4198.30
1905	1149700	1082400	8.87	65.12	4197.60
1906	2278600	1808600	13.25	56.98	4198.30
1907	4285100	3831100	11.54	61.69	4200.50
1908	2312100	1873100	12.06	61.12	4201.00
1909	4480900	4057900	13.41	61.50	4202.60
1910	3416300	3008300	9.11	66.84	4203.90
1911	2320900	1927900	9.95	62.85	4203.20
1912	2450700	2072700	11.31	55.73	4202.70
1913	2822800	2460800	9.93	59.77	4202.90
1914	2950400	2603400	12.27	59.01	4203.60
1915	1630800	1399400	10.79	58.70	4203.20
1916	2434000	2115000	8.92	59.32	4202.90
1917	3039900	2735000	14.12	52.41	4203.40
1918	2921700	2629700	8.09	65.27	4203.50

Table 1. Continued - Water Balance Model Input Data and Historical Annual Peak Elevation.

Water Year	Historical River Inflow (acre-feet)	Modified River Inflow (acre-feet)	Lake Precipitation (inches)	Pan Evaporation (inches)	Annual Peak Lake Level (feet-msl)
1919	1930400	1647400	8.28	67.00	4202.70
1920	2174700	1909700	11.22	57.80	4202.00
1921	3756000	3498000	11.65	59.97	4203.30
1922	4227700	3977700	11.36	63.80	4204.40
1923	3014400	2771400	11.50	55.47	4204.90
1924	2915600	2680600	6.86	64.65	4205.10
1925	2171900	1943900	14.87	56.25	4204.30
1926	1961900	1739900	10.38	61.47	4204.30
1927	1985100	1765100	10.00	58.54	4203.50
1928	2113900	1893900	6.77	65.02	4202.60
1929	1139700	1139700	11.45	57.49	4202.00
1930	3442200	3214200	10.76	61.25	4201.20
1931	879600	811200	6.23	67.88	4200.45
1932	1768900	1406200	11.02	58.82	4199.40
1933	1604900	1243300	7.25	63.54	4198.85
1934	1385800	599600	5.50	76.57	4197.20
1935	998000	799200	9.98	62.07	4196.05
1936	1645600	1468600	10.20	63.08	4195.80
1937	1780900	1399200	11.12	59.80	4196.45
1938	1729100	1486400	10.76	68.02	4196.55
1939	1730700	1122400	10.08	69.25	4196.55
1940	1433600	861000	8.21	73.96	4195.75
1941	1327900	940100	14.11	60.67	4195.65
1942	1900400	1399300	11.47	59.96	4196.55
1943	1797200	1575500	8.42	67.25	4196.20
1944	1351400	1305800	10.82	62.87	4196.45
1945	1470200	1417500	11.35	59.28	4196.40
1946	1842000	1807100	10.14	62.14	4197.15
1947	1676600	1630000	12.43	58.50	4197.20
1948	1964000	1899000	9.49	62.37	4197.75
1949	1907000	1802900	10.82	58.34	4198.25
1950	2745800	2632500	10.43	57.52	4198.80
1951	2687600	2484800	11.38	62.21	4199.90
1952	3184700	2580700	10.80	60.05	4200.95
1953	1962000	1768100	8.91	60.47	4200.55
1954	1001000	955300	6.65	66.10	4199.35
1955	1036400	921800	9.31	58.20	4198.05
1956	1496200	1392800	9.01	60.42	4197.85
1957	1623900	1505500	11.73	57.78	4197.45
1958	1627700	1736600	8.82	64.82	4197.40
1959	970000	935500	8.83	61.52	4196.05
1960	944500	862800	6.56	67.71	4195.30
1961	659200	597600	7.97	67.27	4193.80
1962	1308500	1253300	11.93	61.20	4193.85
1963	992200	921800	8.59	59.11	4192.95
1964	1447600	1392800	10.46	57.79	4194.15
1965	1870500	1505500	12.04	55.05	4194.25
1966	1584200	1584200	6.57	64.26	4195.60
1967	1620500	1620500	11.73	55.87	4195.15
1968	1652300	1652300	12.42	60.04	4195.60
1969	2182700	2182700	10.73	64.46	4197.10
1970	1593500	1593500	10.89	62.85	4196.40
1971	3009700	3009700	13.15	59.53	4198.00
1972	3071700	3071700	10.32	60.66	4199.60
1973	2364000	2364000	14.81	61.44	4200.45
1974	2545300	2545300	9.19	67.71	4201.30
1975	2337200	2337200	13.45	56.38	4201.55
1976	2535700	2535700	10.81	57.13	4202.25
1977	1068900	1068900	9.74	59.23	4200.75
1978	1637700	1637700	12.56	54.93	4200.25
1979	1453700	1453700	9.36	55.77	4199.90
1980	2493400	2493400	13.25	55.54	4200.55
1981	1496600	1496600	9.31	58.67	4200.15
1982	2440000	2440000	17.23	55.24	4200.70
1983	5304300	5304300	17.79	52.54	4204.70
1984	6649300	6649300	16.61	56.19	4209.25
MEAN	2434244	1988118	10.64	61.02	4201.77

for the 1944-1983 calibration period and the 1984 peak are shown on Figure 2. The model computed the 1984 peak to be 4209.03 compared to the measured peak of 4209.25 feet. This is an error of 0.21 feet which is typical of the errors for the calibration period. Table 2 shows the results of the water balance model for a mean annual water budget for the Great Salt Lake.

Table 2. Great Salt Lake Water Budget with Present Level of Water Use and 1851-1983 time base (Acre-Feet).

INFLOW		
	Gaged or correlated streamflow	1,950,000
	Estimated ungaged surface water	150,000
	Estimated ungaged ground water	85,000
	Subtotal	2,185,000
	Precipitation (10.6 inches)	815,000
	Total Supply	3,000,000
OUTFLOW		
	Evaporation (38.9 inches)	3,000,000

The water balance model was developed primarily for determining the effects of water development projects on lake levels and to evaluate proposed lake management alternatives such as the west desert pumping plan. Figure 3 shows the effects of 350,000 acre-feet of depletion compared to the present modified inflows. This amount of depletion is approximately Utah's allocation of the Bear River Compact and would have an effect of lowering the lake as much as 6 feet in dry years and as little as 2 feet in extremely wet years. Figure 4 shows the effects the west desert pumping plan would have on lake levels compared to the present modified hydrograph. The pumping plan would have little effect on low lake levels and can have significant effects reaching nearly 5 feet during high lake levels depending on the rate of inflow. Pumping would have reduced the 1984 peak level by 2 feet which when combined with the causeway breach could have reduced the 1984 peak by 3 feet.

The water balance model was used to estimate where the lake would have been for pristine conditions (adjusted to eliminate man's uses as they have occured over history) and where it might go in the future based on a percentage of average runoff. Figure 5 shows the pristine hydrograph compared to the historical and present modified hydrographs. For pristine conditions the 1984 peak would have been 4215.7 feet compared to 4209.25 which is approximately 6 feet higher. Figure 6 shows where the lake would go based on a percentage of average river inflow. Average inflow of 150 percent would keep the lake near an elevation of 4210 feet. If the inflow would return to average conditions it would take until the year 2000 (15 years) for the lake to return to its long-term average near elevation 4200 feet.

The water balance model is currently being used by the Utah Water Research Laboratory in conjunction with a stochastic flow generation model and a damage simulation model to access the hydrologic effects and economic analysis of lake level control alternatives.

FIGURE 3. Great Salt Lake Hydrographs—Modified and 350000 ac-ft Depletion

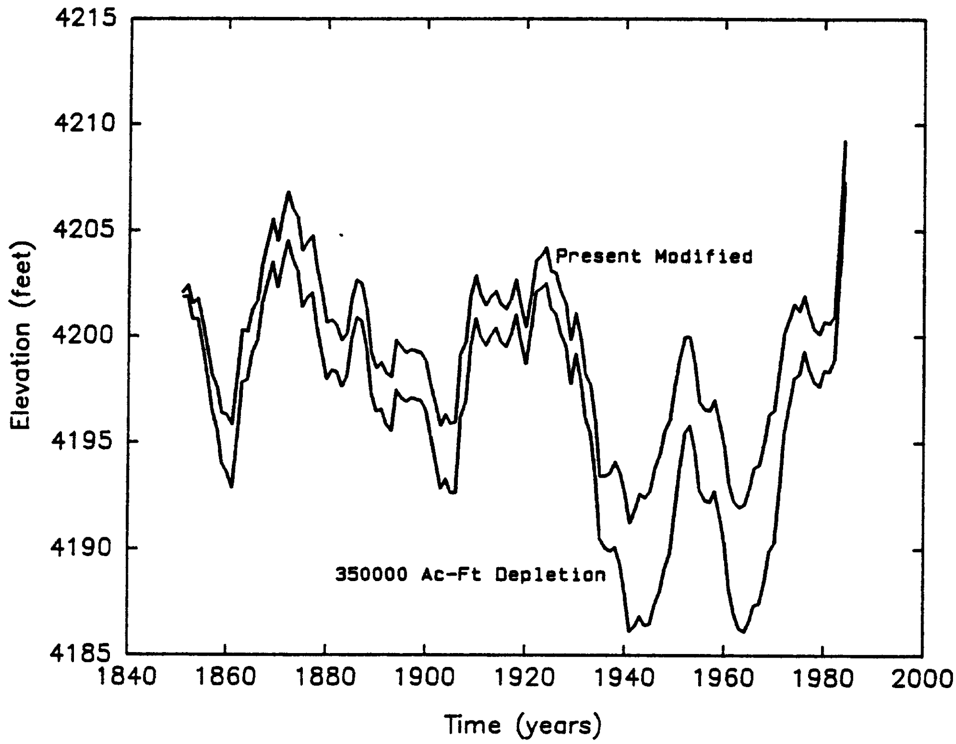


FIGURE 4. Great Salt Lake Hydrographs—Modified and West Desert Pumping

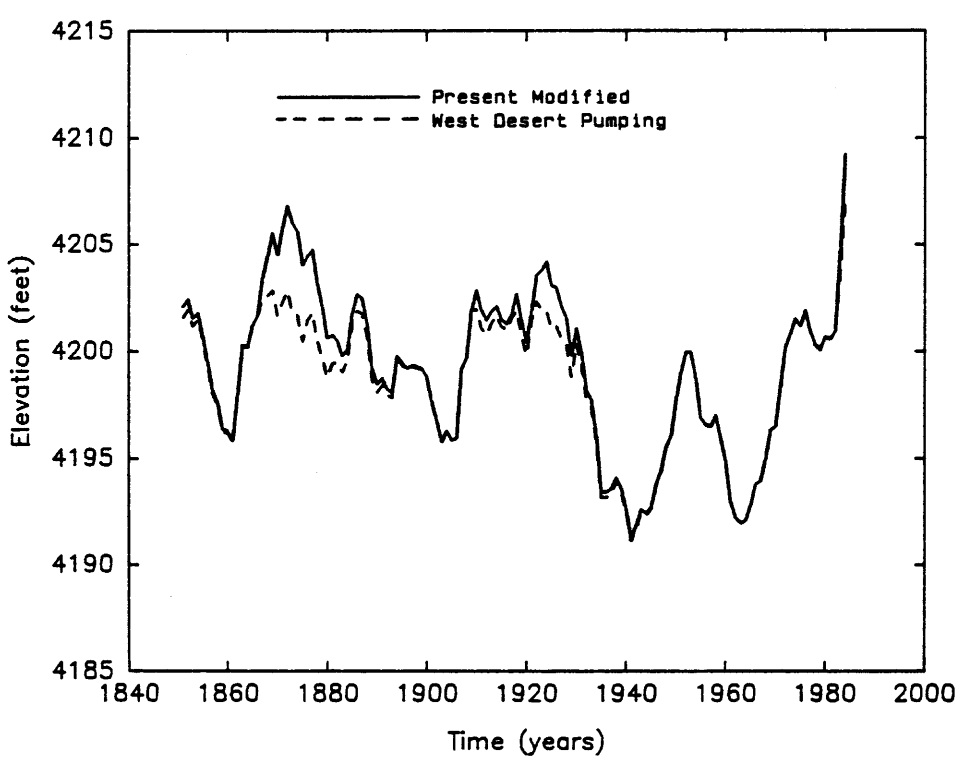


FIGURE 5. Great Salt Lake Hydrograph Peaks—Pristine, Historic, and Modified

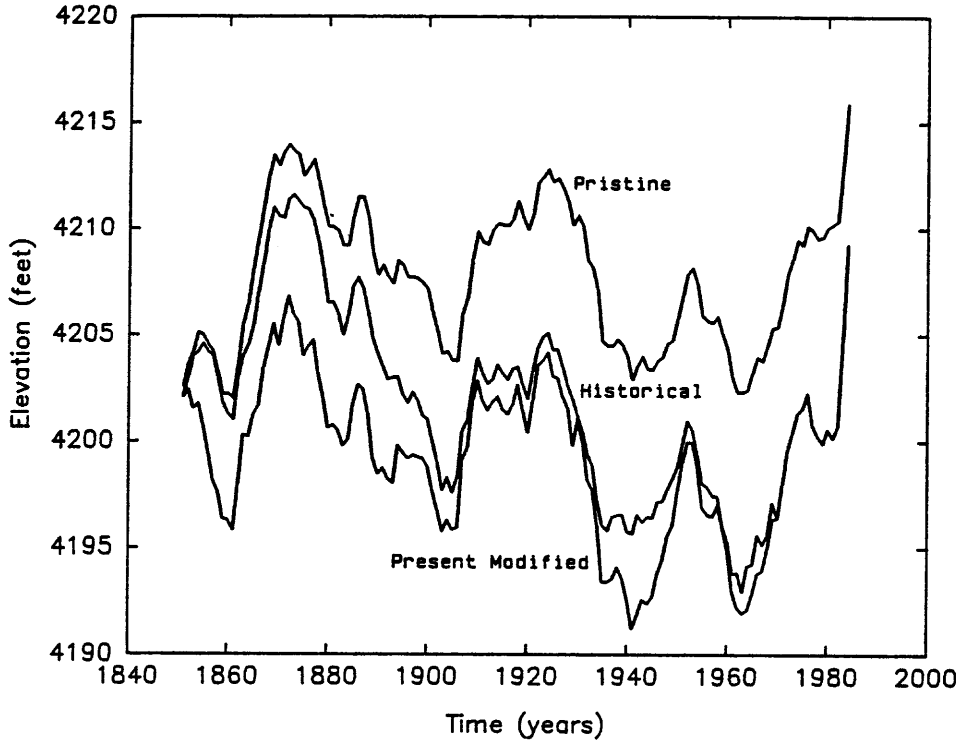
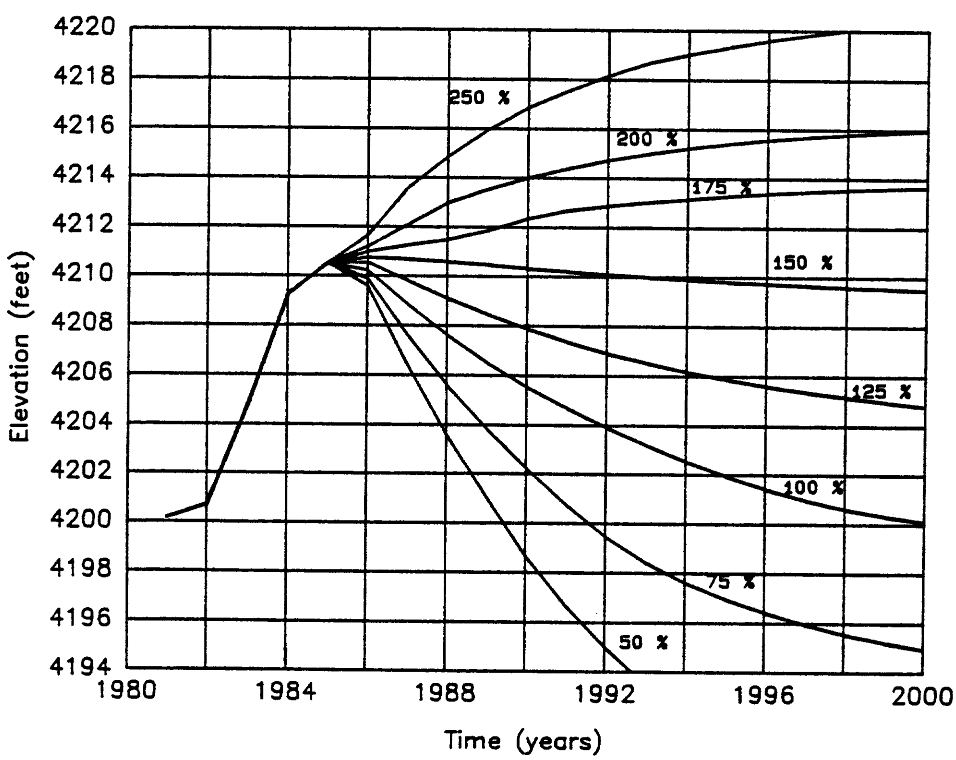


FIGURE 6. Great Salt Lake Projections as a function of Average Runoff



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FORECASTING GREAT SALT LAKE LEVELS[1]

Thomas E. Croley II[2]

The Great Lakes Environmental Research Laboratory (GLERL) developed a semi-automatic software package for making deterministic outlooks of basin moisture storage conditions, basin runoff, net lake supplies, and lake levels six full months into the future for large lakes. We designed the package especially for use on small computers with a standard FORTRAN-77 compiler, 5 MB of disk storage, and a minimum of CPU and memory resources. The package incorporates the GLERL Large Basin Runoff Model (LBRM) applied to each of the 20 to 40 subbasins about a lake and combined to represent the entire basin. Our near real-time data reduction system uses new algorithms to efficiently determine daily areal averages of meteorologic variables over each of the subbasins. The model and the data reduction system are combined into a useful, easy-to-use, semi-automatic package that consists of easily-supported modules for making near real-time forecasts of basin runoff and lake levels. Application of the runoff model and the forecast package to the Great Salt Lake in Utah involves refinement of several model concepts since the hydrology differs from the Laurentian Great Lakes, where the model was developed.

Introduction

Near real-time outlooks of a lake's net basin supply (basin runoff plus lake precipitation minus lake evaporation) enable lake-level forecasts. With the intrinsic memory of large basins and lakes, there is much potential for developing useful short-term operational forecasts in the face of uncertain meteorology. The Great Lakes Environmental Research Laboratory (GLERL) has developed model packages that use near real-time information to establish forecasts which consider both the existing basin storages and anticipated or forecast meteorology. These packages are based on GLERL's Large Basin Runoff Model (LBRM), applied to each of the subbasins about a lake as described in the next section. GLERL has used its LBRM in lake level forecast packages for Lake Superior, developed for the U. S. Army Corps of Engineers (Detroit District), and for Lake Champlain, developed for the National Weather Service Northeast River Forecast Center. Special considerations must be made for application of the model to the Great Salt Lake Basin; these are discussed following the description of the LBRM. Our forecast package enables us to consider the historic record in applying the runoff model, provisional data as it becomes available in making estimates of current basin moisture conditions, and forecasted meteorology in predicting basin runoff, net basin supply, and lake levels. The forecast package is outlined following the discussion of the Great Salt Lake runoff model application.

[1]GLERL Contribution No. 451.

[2]Research Hydrologist, Great Lakes Environmental Research Laboratory, Environmental Research Laboratories, Office of Atmospheric and Oceanic Research, National Oceanic and Atmospheric Administration, U.S. Department of Commerce, 2300 Washtenaw Avenue, Ann Arbor, Michigan 48104.

Runoff Modeling

The GLERL LBRM, pictured schematically in Fig. 1, is an interdependent tank-cascade model which employs analytical solutions of climatological considerations relevant for large watersheds (Croley, 1983a,b). Water enters the snowpack, if present, and some then infiltrates into the upper soil zone based on degree-day determination of snowmelt and net supply; infiltration is taken as instantaneously proportional to the supply rate and to the areal extent of the non-saturated portion of the upper soil zone (partial-area infiltration concept). Percolation from the upper zone enters the lower soil zone and percolation from the lower zone enters the groundwater zone. Lateral flows from these zones of surface runoff, interflow, and groundwater flow, respectively, enter the surface storage zone which represents surface waters that ultimately flow from the basin. All of these flows are taken as instantaneously proportional to their respective storages (linear-reservoir flow concept). Evapotranspiration occurs from the two soil zones as a function of the available water and of the heat available for evaportranspiration; the heat used for evapotranspiration reduces the opportunity for additional evapotranspiration (complementary evapotranspiration and evapotranspiration opportunity concept). The mass balances for snowpack, upper and lower soil storages, groundwater, and surface water use these physically-based concepts, in the cascade of Fig. 1, to form a set of simultaneous ordinary linear differential equations whose joint solution depends upon the relative magnitude of all parameters, inputs, and system states (storages) pictured in Fig. 1. Complete analytical solutions for all possible ranges of values are available (Croley, 1982). Simulation and forecasting require only daily maximum and minimum air temperatures and precipitation data and initial storage values.

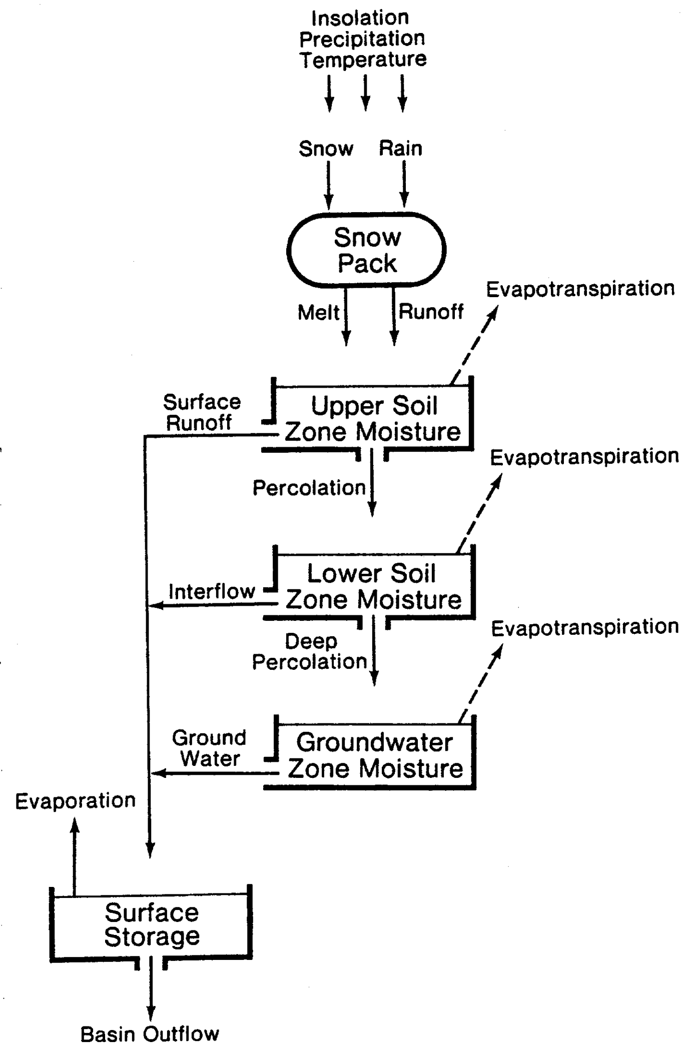

Fig. 1. Runoff Model Schematic

In applying the model to a large drainage basin, the basin is first divided into subbasins draining directly to the lake. About 30 years of meteorological data from all available stations (typically 150-300) about and in the subbasins are combined through Thiessen weighting to produce areally-averaged daily time series of precipitation and minimum and maximum air temperatures for each subbasin. Likewise, all available records from the most-downstream flow gauges are combined to estimate the daily runoff to the lake from each subbasin. Procedures for this climatic data reduction are completely automated; all climatic data can be prepared in little more time than it takes to receive the data. The model then uses the weighted meteorological data in a "distributed-parameter" application to compute model outflows from each of the subbasins; these outflows are combined to produce the entire-basin runoff.

Each application of the LBRM requires determination of the values of the 9 parameters shown in Fig. 1; the parameter calibration is described in detail elsewhere (Croley and Hartmann, 1984). Parameters are determined in a semi-automated systematic search of the parameter space to minimize the sum-of-squared errors between actual outflow volumes and model outflow volumes. The search consists of minimizing this error for each parameter, selected in rotation, until convergence to two significant figures in all parameter values is achieved. The model is used over an initialization period of two or more years prior to the calibration period; the effect of initial conditions (basin storages) diminishes with the length of the simulation. The calibration then is repeated by using long-term averages for the initial conditions at the beginning of the calibration period.

Studies on several of the Great Lake Basins show that the search algorithm does not give unique optimums because of synergistic relationships between parameters. Since parameter error compensation is probably present in the calibrations, the algorithm yields multiple optimums dependent on the starting parameter set. However, frequency dependence of some of the parameters is used to determine their values. For example, parameters governing the slow-responding groundwater storage are determined with a monthly response model applied to monthly data. Likewise, the lower soil zone response may be fairly well-defined by a weekly model applied to weekly data, and the upper soil zone, surface, and snowpack responses may be defined adequately only in a daily model applied to daily data. The model is generally first used in a lumped-parameter application to a monthly entire-basin data set, formed by combining all subbasin data sets and accumulating over the monthly interval; many data measurement errors are effectively filtered as well as the higher-frequency components. The monthly model is then used on subbasin monthly data sets, with the initial parameter values taken from the preceding calibrations. This is repeated for the weekly and daily models, each time better defining the parameter values of preceding calibrations.

The LBRM captures a "realism" in its structure that has several advantages over other models. Basin storages, modeled as "tanks", are automatically removed as respective parameters approach their limits. Thus, the structure of the model changes within a calibration. This is achieved without the use of "threshold" parameters in the model; physical concepts are used which avoid discontinuities in the goodness-of-fit as a function of the parameters. Initialization of the model corresponds to identifying storages from earlier simulations or from field conditions which may be measured; interpretations of a basin's hydrology then can aid in setting both initial and boundary conditions as well as in setting parameter values. Areal measurements of snowpack and soil moisture water equivalents, available from aerial or satellite monitoring, are used in the Lake Superior applications (Gauthier et. al., 1984).

The correlation and root mean square error of the LBRM in example Lake Superior daily applications (Croley and Hartmann, 1984) were 0.92 and 0.25 mm (expressed as a depth over the basin surface) which may be compared to 0.77 and 0.40 mm from a climatic approach that uses mean daily flow as a predicter of basin runoff. Comparisons on other lakes with other models also show the LBRM to be superior (Croley, 1983a,b).

Great Salt Lake Application

The Great Salt Lake lies at approximately 1280 meters above sea level within the lowest portions of the Great Basin and drains an area of approximately 61000 square kilometers (including the lake). Being a terminal lake, there is no outflow to the sea; water may leave the lake only through lake evaporation, groundwater flow, and limited consumptive use. The Wasatch Mountain Range east of the lake reaches elevations in excess of 3350 meters and contributes significantly to the inflow of the lake (about 92% of surface contributions during 1931-1976) mainly through the Bear (59%), Weber (20%), and Jordan (13%) Rivers (Keck and Hassibe, 1983; Arnow, 1984). The Great Salt Lake Desert west of the lake, once flooded by ancient Lake Bonneville, contributes less than 5% of the total runoff to the lake. Normally (1931-1976), about 2.3 billion cubic meters of water enters the lake annually as surface runoff (66%), about 1.1 billion cubic meters enters as lake precipitation (31%), about 0.1 billion cubic meters enters as groundwater (3%), and about 3.5 billion cubic meters leaves the lake as evaporation (100%). More recently, these numbers were 2.6, 1.4, 0.1, and 4.1 respectively. Additionally, consumption in 1976 was less than 0.1 billion cubic meters. Minerals dissolved in the surface runoff remain in the lake as almost all water outflow is through lake evaporation, making the lake one of the saltiest bodies of water in the world. The imbalance between inflow and outflow results in changes to the lake's volume; due to the small relief of the lakebed, a small change in water depth results in a large change in the areal extent of the lake. Due to its shallowness, the surface-water temperature of the lake varies considerably from about -7 degrees Celsius in January to about 27 degrees Celsius in early August. Due to its high salt content, the lake does not freeze and thus moderates the influence of the cold air on its environs. The lake's relative warmth, on the other hand, contributes moisture resulting in heavy snowfalls about the lake. As the lake storage increases, surface area increases, salinity decreases, heat storage increases, and the net effect is an increase in lake evaporation, although the evaporation process is poorly understood at present. The lake is divided into northern and southern sections by a railway causeway, completed in 1959; limitations on flow through the causeway results in a difference, between the two sections, in water temperature, salinity, and water level.

There are essential differences in the hydrology and basin dynamics between the Great Salt Lake and the Laurentian Great Lakes; the LBRM must be modified in its construction and use, to recognize these differences. There is a dynamic interaction between the Great Salt Lake and its basin not found to the same extent in the Laurentian Great Lakes. This interaction is related to changes in the lake area (and consequently contributing basin surface area) that accompany changes in lake level. Also, the complementary relationship between evapotranspiration and the heat available for evapotranspiration, which is used in the Laurentian Great Lakes, does not consider lake evaporation; that is, lake evaporation is estimated separately from basin evapotranspiration. While this arrangement works well in the humid mid-west, it is likely that lake evaporation and basin evapotranspiration for the Great Salt Lake must be considered jointly in the complementary relationship. Lake evaporation also depends on the lake area, which must be recognized in the joint heat balance for the lake and basin. Furthermore, the salinity of the Great Salt Lake is highly dependent on the water content of the lake, and the relationship of salinity to lake evaporation is not clearly understood.

The groundwater flow to the Great Salt Lake is estimated at about 93 million cubic meters annually, which represents about 3% of the surface runoff to the lake. Depending on the seasonal nature of the NET groundwater flux and on the magnitude of errors associated with the estimation of lake evaporation and net groundwater flux in a water-budget over the lake, groundwater considerations may be necessary for certain times of the year. The aquifers in the vicinity of the Great Salt Lake are complex and direct estimation of the seasonal net groundwater flux to the lake is unlikely. However, it may be possible, if necessary, to index the net flux to the model's groundwater storage, as is being considered for Lake Champlain.

Some components in the conceptual model (such as linear reservoirs) are more likely to adequately represent their processes in the real world than others (such as degree-day melting or complementary evapotranspiration). These concepts were established in research on the Laurentian Great Lakes and can be modified for application to mountainous and less humid areas. Parameter estimation techniques that properly weight the more accurate parts of the model could also improve parameter estimates. Model performance is also expected to improve with the use of additional data that may be available from upgraded meteorological stations (e.g., relative humidity, wind speed and direction, soil temperatures, and solar radiation), by incorporating more-sophisticated concepts within the model.

To make timely forecasts, a responsive data collection and reporting system must exist. In the Lake Superior Basin, our first concern was the requisite density of the data network required for reliable modeling. Special studies of network size and station locations for both meteorologic and hydrologic gauges, carried out in cooperation with the U.S. Army Corps of Engineers, Detroit District, indicate that one station per 4300 to 6500 square kilometers is suitable for estimation of weekly runoff. A special study also looked at the effect of data availability on our forecasts. We found that modeling explains 78% of the variance of basin outflow on Lake Superior with no delay in processing data, 70% with a 1-week delay, and 66% with a 2-week delay. With a 10-week delay, forecasts are no better than using data available with a 2-year delay. We are designing now for a 1-week data lag.

Other data requirements and modeling improvements remain. The modeling of basin evapotranspiration should improve if we can actually measure the total heat available (rather than relying on a simple function of air temperature). Improved lake evaporation models are expected that can use near-real-time lake surface temperatures. Also, the LBRM was designed to describe only the natural basin response to meteorological conditions. Thus, we require information about regulation decision rules or the ability to monitor regulated flows in a timely fashion.

Forecast Package

Forecasts are integrations of modeling and near real-time data handling. Our forecast package, outlined schematically in Fig. 2, consists of 6 modules for: 1) preparing climatic subbasin files of areally-averaged historic meteorologic and hydrologic data, 2) calibrating and applying the runoff model for each subbasin, 3) preparing initial data bases for the forecast modules, including climatic hydrometeorologic quantiles to aid in selecting a forecast meteorologic scenario and to provide perspective for the forecasts, 4) updating provisional data bases with near real-time meteorologic data, 5) selecting a forecast meteorologic scenario, and 6) transforming forecast meteorology into forecast basin runoff, net basin supply, and lake levels. The first three modules are used infrequently as new climatologic data becomes available every one or two years; they are implemented for a mini or mainframe computer, since the computations are extensive. The last three modules are used often as provisional data becomes available and as forecasts are desired. The last three modules operate in a highly efficient manner and are implemented for speed and a minimum of CPU time, memory, and disk storage; they thus can be used efficiently on either large or small computers, and are transferred to the agencies responsible for forecasts on a given lake. These modules are described sequentially in the following paragraphs.

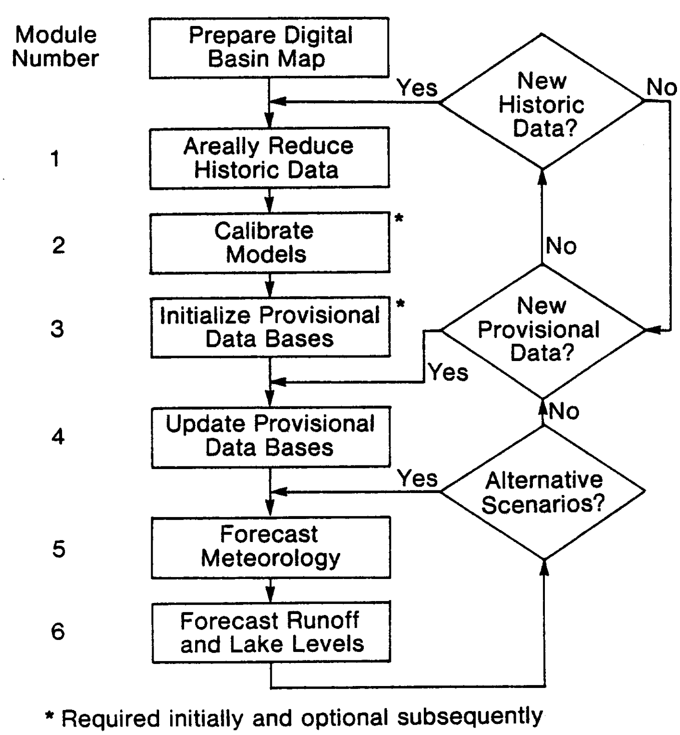

Fig. 2. Forecast Package Schematic

Before using the first module in Fig. 2, a "digital map" of all subbasins must be prepared for use in the ensuing computations. A digital map is a two-dimensional array representing the areal extent of the subbasins to the nearest square kilometer. For the first module, meteorologic station files are assembled for all stations about and in the subbasins, and hydrologic station files are assembled for all "most-downstream" flow stations in the subbasins. Each meteorologic station file contains daily values of minimum air temperature, maximum air temperature, and precipitation for the length of the historic record as well as the location of the station. The files are ordered by distance from the total basin and the nearest 150-300 stations are identified for subsequent use. For each subbasin, for each day, all stations reporting on that day are considered to compute Thiessen weights for the resulting networks for each of minimum air temperature, maximum air temperature, and precipitation if those networks have not already been considered in previous days. We use a new algorithm (Croley and Hartmann, 1985) for determining weights that is especially advantageous for high-resolution maps and varying high-density data-observation networks; it enables fast computation of daily areally averaged subbasin meteorology. The hydrologic station files are then used to estimate the historic subbasin outflows. Flows from more than one gauge in a subbasin are combined by aggregating and extrapolating for ungauged areas.

Following reduction of the historic data, the LBRM is calibrated for each subbasin (the second module in Fig. 2) by following the procedures described above. The third module in Fig. 2 combines all subbasin data and estimates quantiles of basin storages, runoff, and net basin supply (to provide a perspective by which to interpret subsequent forecasts) and quantiles of daily average air temperature and precipitation (used in selecting candidate forecast meteorological scenarios corresponding to the National Weather Service monthly and seasonal outlooks). Exceedance probabilities associated with the latter quantiles are then estimated for each period in each year; e. g., the 1964 January exceedance probability corresponding to the 50% January precipitation quantile is estimated by the relative number of days in January 1964 that precipitation exceeds this quantile. Finally, this module reformats and initializes data bases for use with the forecast modules.

The fourth module in Fig. 2 accepts new provisional data and adds it to the provisional data bases. Provisional data are temperatures and precipitation values that are reported in near real-time as available. Effective operational forecasts require reception and processing of meteorological data from all stations (typically 20 - 50) in a provisional measurement network within a week after measurement. Provisional data are contained in individual station files; each file contains the latest daily values of minimum and maximum air temperature and precipitation since the last time that the data bases were updated. The station files may also contain new, corrected, or changed data values for any days prior to the last update; sequential or chronological order is unnecessary. All provisional station files are checked for format and range errors and are incorporated into a provisional data base; only new or changed data are incorporated and the corresponding dates are "flagged". The provisional data base is then scanned and only flagged dates are considered to compute the daily Thiessen-weighted provisional subbasin areal-averages which overwrite old values in the provisional subbasin data files or are appended to them; the flags are then removed. The subbasin provisional data files are then used with the LBRM to update subbasin storage conditions; the updates are made only from the point of the earliest change or addition of provisional data to the end of the provisional data. Field measurements of basin moisture conditions, available from snow course, aerial, or satellite monitoring, may be incorporated for the LBRM in this module; model-generated storages are replaced by known field conditions on the date corresponding to their measurement. Thus, current subbasin storages are estimated with the model as applied to the most recent provisional data and available field measurements, and they serve as initial conditions for a forecast.

Basin runoff is forecast by applying the LBRM for each subbasin to a forecast meteorology scenario. The selection of this scenario is important and difficult; the resulting forecast of runoff will be no better than the forecast of air temperatures and precipitation used in the simulation. Several methods are available; the one used here preserves the spatial and temporal interdependencies of all meteorologic variables and recognizes the limited expertise available in forecasting meteorology. The U. S. National Weather Service provides monthly and seasonal weather outlooks bi-monthly for the North American Continent (NWS, 1984) which consist of maps of air temperature and precipitation probabilities for the coming month and 90-d season. Module 5 in Fig. 2 uses these outlooks to construct a forecast meteorologic scenario; the meteorologic quantile exceedance probability tables (compiled in module 3) are scanned to identify the one or more years of the historical record which best match the forecasts over the forecast period of interest. For each subbasin, the historical daily values of areally-averaged minimum and maximum air temperatures, precipitation, and lake evaporation corresponding to the forecast period of interest are then taken from the identified year(s) of record and used as the forecast meteorologic scenario(s).

The sixth module in Fig. 2 accepts the forecast meteorologic scenario from module 5 and uses it with the LBRM for each subbasin to automatically simulate the resulting basin runoff; the forecast basin runoff is then predicated on the forecast meteorology and the basin storages that exist at the beginning of the forecast period (end of the provisional data period as computed in module 4). The subbasin runoff and storages are each aggregated over the entire basin and the total basin runoff, lake precipitation, and lake evaporation are combined to forecast net basin supply for the lake. Net basin supply is used in conjunction with hydraulic routing models to calculate forecasted lake levels. Various forecast statistics are computed, the forecast entire-basin storage conditions are plotted with historical quantiles for perspective, and the forecast meteorology is (optionally) plotted. A partial example is pictured in Fig. 3. An alternative use of the forecast package of Fig. 2 allows the generation of Extended Streamflow Prediction (ESP) forecasts. Module 5 in Fig. 2 consists simply of selecting a year of record for the meteorological forecast, and modules 5 and 6 are repeated for every year of record. The resulting "runoff and lake level scenarios" are analyzed statistically to determine, e. g., the forecast 5%-, 50%-, and 95%-exceedance lake levels.

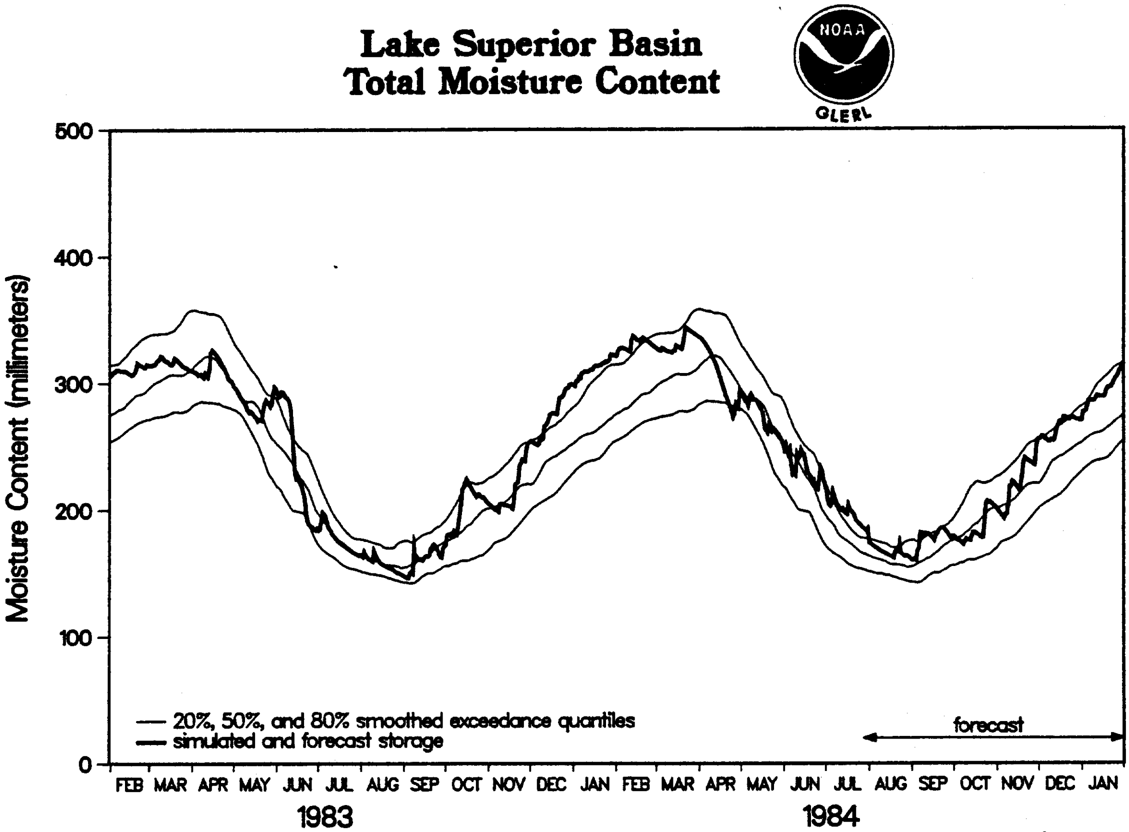

Fig. 3. Example Forecast for Lake Superior (Total Moisture)

The six modules of Fig. 2 are reused as required for successive product updates, provisional data updates, and forecasts as illustrated thereon. After a forecast is made, modules 5 and 6 are re-executed if there are alternative meteorological scenarios that are desired for consideration. As new provisional data becomes available in near real-time, modules 4-6 are re-executed to include the data and make revised forecasts. As data-collection agencies revise their provisional data and make it available to add to the historic records, the entire package of Fig. 2 may be re-executed to revise the historic data bases, recalibrate the runoff models on each subbasin, and reinitialize the provisional data bases. Modules 1-3 are executed infrequently (every 1-2 years) as historic data is available in machine-readable form; modules 4-6 are executed frequently (every day, week, or month) as provisional data is received in near real-time and as forecasts are desired. Modules 4-6 are encoded to run conveniently on small computers with 5 MB of disk storage; the computer programs in these modules are configured to use a minimum of memory but sufficient memory should be additionally available to contain the digital map of the basin.

Conclusions

The Great Lakes Environmental Research Laboratory has developed a Large Basin Runoff Model for simulations of net basin supply to a large lake. It uses climatic considerations especially relevant for large basins and is applied in a distributed-parameter fashion. We have automated all calibration, climatic data reduction, provisional data reduction, and forecast procedures to enable us to apply the model to large lake basins. The model and forecast procedures have been applied in near real-time to several of the Laurentian Great Lakes and to Lake Champlain. Model modifications are necessary for applications to the Great Salt Lake; these are detailed herein. Potential lake level forecasts in the face of uncertain meteorology are enabled due to the intrinsic memory of large basins and lakes.

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SOME RELATIONSHIPS OF PRECIPITATION AND GREAT SALT LAKE ELEVATIONS, PAST, PRESENT, AND NEAR FUTURE

GERALD WILLIAMS, NATIONAL WEATHER SERVICE, HYDROLOGIST IN CHARGE
COLORADO BASIN RIVER FORECAST CENTER, SALT LAKE CITY, UTAH

INTRODUCTION

The Great Salt Lake is a complex terminal lake, and a reduction in its size is governed only by evaporation in excess of inflow. Evaporation volumes are controlled primarily by atmospheric conditions and lake size. Other secondary influences include salinity and water temperature.

Inflows are primarily from surface runoff and precipitation on the lake. Groundwater inflow is generally a minor contribution averaging less than 1% of the total flow.

BRIEF CLIMATOLOGICAL AND HYDROLOGICAL ANALYSIS

Since 1847 the elevation of the Great Salt Lake has ranged between a low of 4191.35 feet in 1963 and a high of 4211.6 feet in 1873. At the highest level the lake contained a volume near 30 million acre-feet, while at the lowest level the lake contained a volume of less than 9 million acre-feet. Since 1847 the surface area has ranged from near 700,000 acres with 25-30% salinity to near 1,500,000 acres with approximately 12-14% salinity. Presently salinity is near 5% in the south arm but total lake salinity would most likely be greater than 12% if there had been no railroad causeway separating the north end and south end.

Water-year precipitation falling on the lake averages between 10 and 11 inches. The precipitation for the Salt Lake Airport site averages about 15.5 inches and data from that site are used for a long-term comparison to Great Salt Lake elevation. (Table 1 and Figure 1).

Precipitation at 10 representative index stations within the lake's drainage basin averages between 18 and 19 inches. Amounts range from less than eight inches west of the lake to over 50 inches in the higher mountains east of the lake. The volume of evaporation over the lake is somewhat more variable than the depth of evaporation. The depth of evaporation ranges between 35 inches and 50 inches. A USGS study using data from 1931 through 1976 showed evaporation volume ranged from 2.1 million acre-feet to 3.9 million acre-feet, with an average of 2.9 million acre-feet and an average evaporation depth of 45 inches. Generally the highest volume evaporation would be in years when the lake was high, such as 1873 and 1983. At higher lake levels, potentially more evaporation volume would occur due to reduced salinity and greater surface area. However, because climatological conditions vary from year to year, the

Table 1. GREAT SALT LAKE DATA (USGS and NWS)

YEAR	MAXIMUM YEARLY STAGE FEET	MAXIMUM YEARLY VOLUME 1000 AF	MINIMUM YEARLY STAGE FEET	MINIMUM YEARLY VOLUME 1000 AF	WATER YEAR PRECIP INCHES	YEARLY STAGE RISE FEET	YEARLY STAGE FALL FEET	YEARLY VOLUME RISE 1000 AF	YEARLY VOLUME FALL 1000 AF	SEASONAL VOLUME CHANGE 1000 AF
1878	4209.40	26710	4207.80	24224	22.16	0.20	-1.60	303	-2486	-1515
1879	4208.10	24639	4205.60	21729	10.22	0.30	-2.50	415	-2910	-2071
1880	4206.50	22737	4204.80	20756	12.49	0.90	-1.70	1008	-1981	-1902
1881	4206.50	22737	4205.10	21117	15.48	1.70	-1.40	1981	-1620	0
1882	4206.00	22200	4204.50	20400	16.50	0.90	-1.50	1083	-1800	-537
1883	4205.00	21000	4203.50	19237	13.37	0.50	-1.50	600	-1763	-1200
1884	4205.90	22083	4205.30	21356	19.71	2.40	-0.60	2846	-727	1083
1885	4207.30	23646	4206.20	22427	18.06	2.00	-1.10	2290	-1219	1563
1886	4207.70	24097	4206.50	22737	18.46	1.50	-1.20	1670	-1360	451
1887	4207.20	23531	4205.60	21729	14.60	0.70	-1.60	794	-1802	-566
1888	4206.10	22312	4203.90	19684	10.71	0.50	-2.20	583	-2628	-1219
1889	4204.80	20756	4202.10	17613	14.21	0.90	-2.70	1072	-3143	-1556
1890	4204.10	19913	4202.70	18350	16.73	2.00	-1.40	2300	-1563	-843
1891	4203.50	19237	4202.30	17846	13.43	0.80	-1.20	887	-1391	-676
1892	4203.00	18700	4201.70	17159	13.78	0.70	-1.30	854	-1541	-537
1893	4203.00	18700	4201.30	16716	17.43	1.30	-1.70	1541	-1984	0
1894	4203.00	18700	4201.60	17045	17.27	1.70	-1.40	1984	-1655	0
1896	4201.80	17268	4200.40	15798	16.75	0.20	-1.40	223	-1470	-1432
1897	4202.30	17846	4200.50	15887	16.81	1.90	-1.80	2048	-1959	578
1898	4201.90	17384	4199.80	15197	15.81	1.40	-2.10	1497	-2187	-462
1899	4201.20	16611	4199.90	15298	17.39	1.40	-1.30	1414	-1313	-773
1903	4197.70	13249	4196.10	11879	14.16	99.99	-1.60	99999	-1370	-3362
1904	4198.60	14026	4196.90	12613	17.76	2.50	-1.70	2147	-1413	777
1905	4197.60	13166	4196.00	11800	14.51	0.70	-1.60	553	-1366	-860
1906	4198.30	13757	4197.10	12783	19.53	2.30	-1.20	1957	-974	591
1907	4200.50	15887	4199.80	15197	18.90	3.40	-0.70	3104	-690	2130
1908	4201.00	16400	4199.90	15298	20.26	1.20	-1.10	1203	-1102	513
1909	4202.60	18229	4200.00	15400	20.05	2.70	-2.60	2931	-2829	1829
1910	4203.90	19684	4202.00	17500	10.55	3.90	-1.90	4284	-2184	1455
1911	4203.10	18812	4201.60	17045	15.65	1.10	-1.50	1312	-1767	-872
1912	4202.70	18350	4201.80	17268	17.99	1.10	-0.90	1305	-1082	-462
1913	4202.90	18583	4201.70	17159	18.29	1.10	-1.20	1315	-1424	233
1914	4203.60	19345	4202.40	17966	17.19	1.90	-1.20	2186	-1379	762
1915	4203.20	18927	4201.40	16826	14.53	0.80	-1.80	961	-2101	-418
1916	4202.90	18583	4201.30	16716	13.85	1.50	-1.60	1757	-1867	-344
1917	4203.40	19150	4202.30	17846	18.46	2.10	-1.10	2434	-1304	567
1918	4203.50	19237	4201.90	17384	14.41	1.20	-1.60	1391	-1853	87
1919	4202.70	18350	4200.70	16089	12.13	0.80	-2.00	966	-2261	-887
1920	4202.00	17500	4200.70	16089	19.42	1.30	-1.30	1411	-1411	-850
1921	4203.30	19037	4201.90	17384	16.68	2.60	-1.40	2948	-1653	1537
1922	4204.40	20266	4202.90	18583	19.40	2.50	-1.50	2882	-1683	1229
1923	4204.90	20878	4203.70	19459	21.72	2.00	-1.20	2295	-1419	612
1924	4205.10	21117	4203.00	18700	11.25	1.40	-2.10	1658	-2417	239
1925	4204.30	20146	4203.30	19037	20.97	1.30	-1.00	1446	-1109	-971
1926	4204.30	20146	4202.40	17966	14.54	1.00	-1.90	1109	-2180	0
1927	4203.50	19237	4201.80	17268	18.90	1.10	-1.70	1271	-1969	-909
1928	4202.60	18229	4201.00	16400	13.69	0.80	-1.60	961	-1829	-1008
1929	4202.00	17500	4200.25	15648	17.17	1.00	-1.75	1100	-1852	-729
1930	4201.20	16611	4199.70	15099	14.23	0.95	-1.50	963	-1512	-889
1931	4200.45	15843	4197.90	13414	9.27	0.75	-2.55	744	-2429	-769
1932	4199.40	14786	4197.75	13289	14.54	1.50	-1.65	1372	-1497	-1057
1933	4198.85	14257	4196.75	12483	11.28	1.10	-2.10	968	-1774	-530
1934	4197.20	12867	4194.80	10958	8.16	0.45	-2.40	385	-1909	-1390
1935	4196.75	12483	4193.85	10294	13.65	1.95	-2.90	1525	-2189	-385
1936	4195.75	11605	4194.35	10644	13.37	1.90	-1.40	1312	-962	-878
1937	4196.45	12199	4195.30	11308	13.42	2.10	-1.15	1555	-891	594
1938	4196.55	12306	4194.90	11029	12.87	1.25	-1.65	998	-1277	108
1939	4196.55	12306	4194.65	10854	12.00	1.65	-1.90	1277	-1453	0
1940	4195.75	11605	4193.70	10189	11.34	1.10	-2.05	752	-1416	-701
1941	4195.65	11531	4194.05	10434	18.17	1.95	-1.60	1342	-1097	-74
1942	4196.55	12306	4194.65	10854	15.49	2.50	-1.90	1872	-1453	775
1943	4196.20	11963	4194.60	10818	12.14	1.55	-1.60	1110	-1145	-343
1944	4196.45	12199	4194.85	10994	18.85	1.85	-1.60	1381	-1205	236
1945	4196.40	12135	4195.20	11239	16.04	1.55	-1.20	1142	-896	-64
1946	4197.15	12825	4195.40	11379	12.35	1.95	-1.75	1586	-1446	690
1947	4197.20	12867	4196.25	12005	18.83	1.80	-0.95	1488	-863	42
1948	4197.75	13289	4196.15	11921	14.36	1.50	-1.60	1285	-1368	422
1949	4198.25	13714	4196.70	12440	16.83	2.10	-1.55	1793	-1274	425
1950	4198.80	14209	4197.55	13127	15.50	2.10	-1.25	1769	-1083	495
1951	4199.90	15298	4198.35	13802	14.18	2.35	-1.55	2172	-1496	1089
1952	4200.95	16347	4199.35	14736	19.29	2.60	-1.60	2545	-1611	1049
1953	4200.55	15937	4198.75	14164	12.37	1.20	-1.80	1201	-1773	-411
1954	4199.35	14736	4197.15	12825	11.78	0.60	-2.20	572	-1911	-1201
1955	4198.05	13542	4196.40	12135	12.24	0.90	-1.65	717	-1407	-1195
1956	4197.85	13372	4195.90	11720	12.53	1.45	-1.95	1237	-1652	-170
1957	4197.45	13059	4195.85	11681	18.77	1.55	-1.60	1339	-1378	-313
1958	4197.40	13031	4195.25	11274	12.81	1.55	-2.15	1350	-1758	-28
1959	4196.05	11840	4194.30	10608	14.12	0.80	-1.75	566	-1232	-1192
1960	4195.30	11308	4193.20	9829	10.43	1.00	-2.10	700	-1479	-532
1961	4193.80	10258	4191.60	8903	11.43	0.60	-2.20	429	-1355	-1050
1962	4193.85	10294	4192.15	9214	16.88	2.25	-1.70	1391	-1080	36
1963	4192.95	9668	4191.35	8766	12.34	0.80	-1.60	454	-903	-626
1964	4194.15	10504	4192.40	9356	15.58	2.80	-1.75	1738	-1148	836
1965	4194.25	10574	4193.75	10224	20.79	1.85	-0.50	1218	-350	70
* 1966	4195.35	11344	4193.15	9796	9.53	1.60	-2.20	1120	-1548	770
* 1967	4194.90	11029	4193.65	10154	16.35	1.75	-1.25	1233	-876	-315
* 1968	4195.30	11308	4194.00	10400	18.84	1.65	-1.30	1155	-908	279
* 1969	4196.70	12440	4194.80	10958	16.75	2.70	-1.90	2040	-1482	1132
* 1970	4196.00	11800	4194.55	10784	17.76	1.20	-1.45	842	-1016	-640
* 1971	4197.50	13087	4196.50	12262	19.86	2.95	-1.00	2303	-825	1287
* 1972	4198.95	14352	4197.50	13087	14.03	2.45	-1.45	2090	-1265	1265
* 1973	4199.80	15197	4198.55	13982	22.26	2.30	-1.25	2110	-1216	845
* 1974	4200.55	15937	4198.65	14073	15.64	2.00	-1.90	1955	-1864	740
* 1975	4200.70	16089	4199.50	14900	17.54	2.05	-1.20	2017	-1189	153
* 1976	4201.50	16937	4199.80	15197	16.31	2.00	-1.70	2037	-1740	848
* 1977	4200.25	15648	4198.30	13757	14.90	0.45	-1.95	451	-1891	-1289
* 1978	4199.60	14998	4197.95	13457	19.23	1.30	-1.65	1241	-1541	-650
* 1979	4199.35	14736	4197.15	12825	8.19	1.40	-2.20	1279	-1911	-262
* 1980	4199.70	15099	4198.50	13937	16.73	2.55	-1.20	2274	-1162	363
* 1981	4199.60	14998	4198.00	13500	13.04	1.10	-1.60	1061	-1498	-101
* 1982	4200.00	15400	4199.35	14736	25.15	2.00	-0.65	1900	-664	402
* 1983	4204.30	20146	4204.05	19857	20.58	4.95	-0.25	5410	-290	4746
* 1984	4208.55	25390	4207.85	24292	23.82	4.50	-0.70	5534	-1098	5244
AVERAGES FROM THE PERIOD OF RECORD ABOVE										
	4200.53	16293	4198.95	14772	15.71	1.61	-1.58	1535	-1521	-28
5 YEAR AVERAGE										
	4202.43	18207	4201.55	17264	19.86	3.02	-0.88	3236	-942	2131

Footnotes.
All changes in volume and stage are calculated from the yearly peak and low in stage height regardless of when it occurred.
Seasonal volume change is calculated using consecutive yearly peaks.
Record was adjusted for the causeway '*'.
99999 represents a discontinuity in the lake record.
The 1984 volume fall includes the causeway breach and thus is not solely due to the effect of evaporation.

HISTORICAL GREAT SALT LAKE HYDROGRAPH
VS. PRECIPITATION AT SLC WSFO FOR VARIOUS INTERVALS

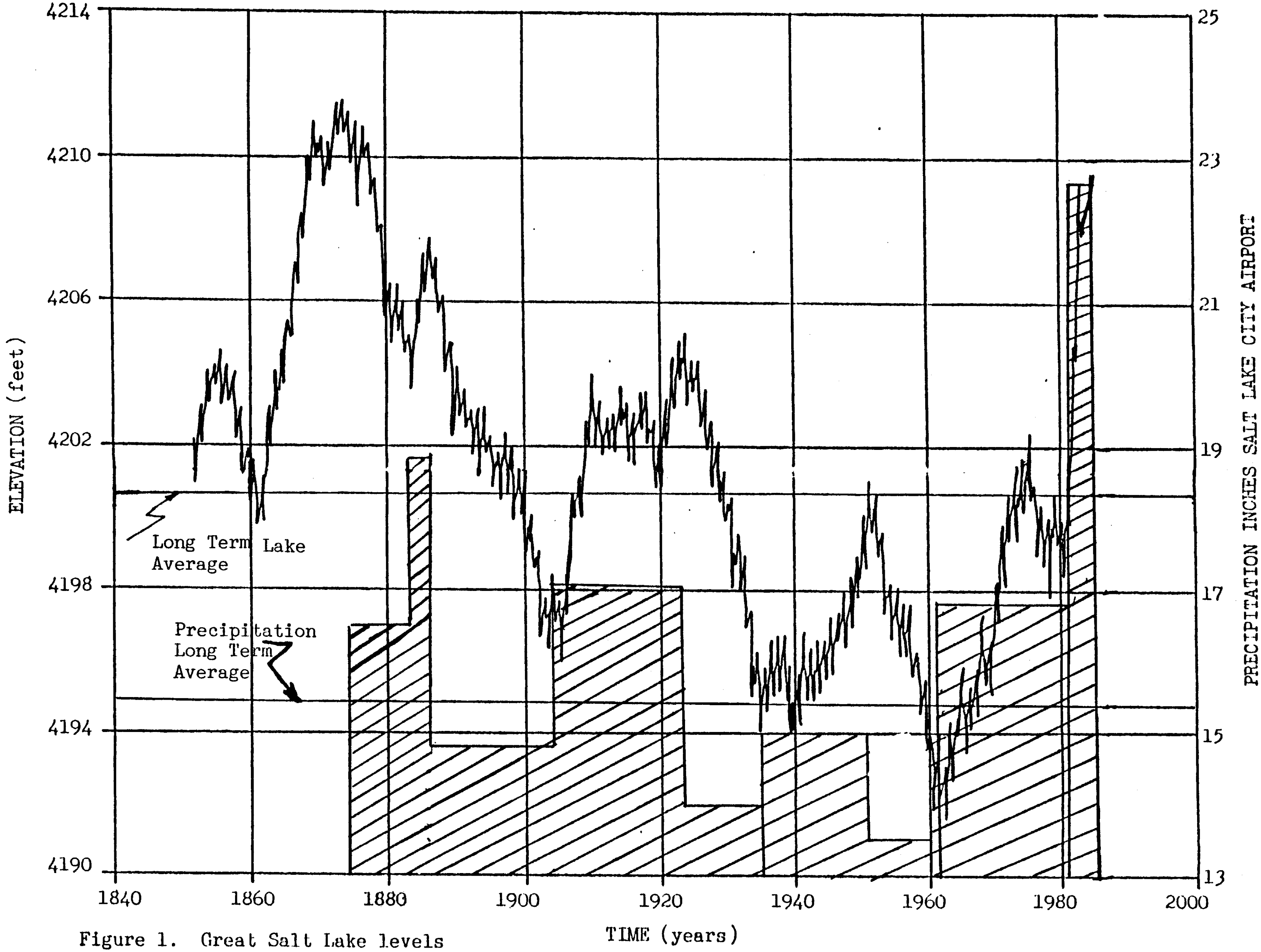

Figure 1. Great Salt Lake levels vs. Precipitation average for selected time intervals at SLC WSFO.

depth of evaporation at a given level may not always be as great as in a year when the lake is at lower levels.

PRECIPITATION ANALYSIS IN CONJUNCTION WITH EXPECTED LEVELS.

Precipitation is the dominant factor regulating inflow to the lake and is the only long-term source of all inflow. Comparisons are presented using the Salt Lake Airport data (which has a long stable record and is easily obtainable), and 10 representative index stations over the basin to provide better hydrologic representation. These analyses were made for two different purposes: 1) to show a relationship of precipitation to lake levels (Figure 2), and 2) to show probability of certain amounts of basin precipitation occurring (Figure 3). Figure 2 was developed through applying a water budget model from above the 4200-foot level and using a precipitation analysis for elevations less than 4200 feet MSL.

To provide a representative precipitation data base, precipitation stations within the basin were analyzed during the 1939-1984 period using October through May precipitation. Specific stations were selected for this period because, 1) a high quality continuous record existed, 2) the period was representative of the precipitation on the lake as well as over the basin which produces surface runoff, and 4) the stations correlated well with surface runoff from the major tributaries and were used in modelling runoff from tributaries to the Great Salt Lake.

PROBABILITY RELATIONSHIPS

To determine likelihood of achieving various lake levels, it seems appropriate to use sound hydrologic principles. The obvious factor appears to be precipitation and corresponding likelihood of occurrences and subsequent effects on the Great Salt Lake.

Figure 3 shows the relationship of precipitation amounts averaged for the 10 stations and the probability of occurrence of various amounts for the October through May period. This is the period most closely correlated with the inflow producing the early summer lake peak level. This figure can also be used to estimate the exceedance probability of total precipitation at any date from October 1 through May 31. To calculate this value, precipitation from October 1 to date must be known and the contingent percentages must be determined subsequent to that date. For example, on March 1, 1985, precipitation October through February is summed and added to contingent values through May. For example, if 1984-85 October through February data (which was near 115%) are used and average amounts are assumed from March through May, a value of near 110% of normal would occur for the entire period. This gives us the most likely October through May amount, 110%, which is the median for an equal probability for seasonal amounts, or is the 50% exceedance value of March 1, 1985. Values for December 1, 1985 are very similiar as shown in figure 3.

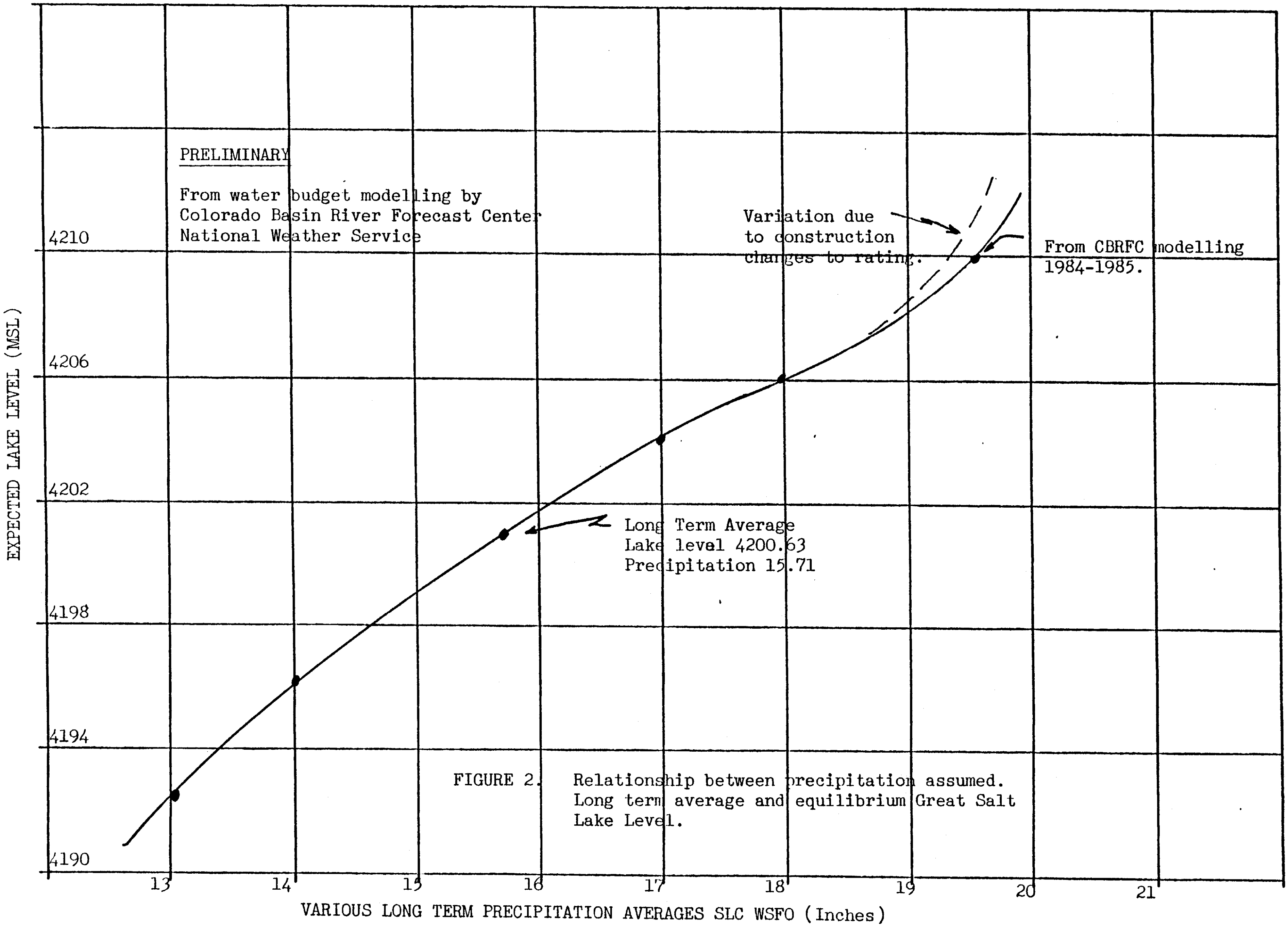

FIGURE 2. Relationship between precipitation assumed. Long term average and equilibrium Great Salt Lake Level.

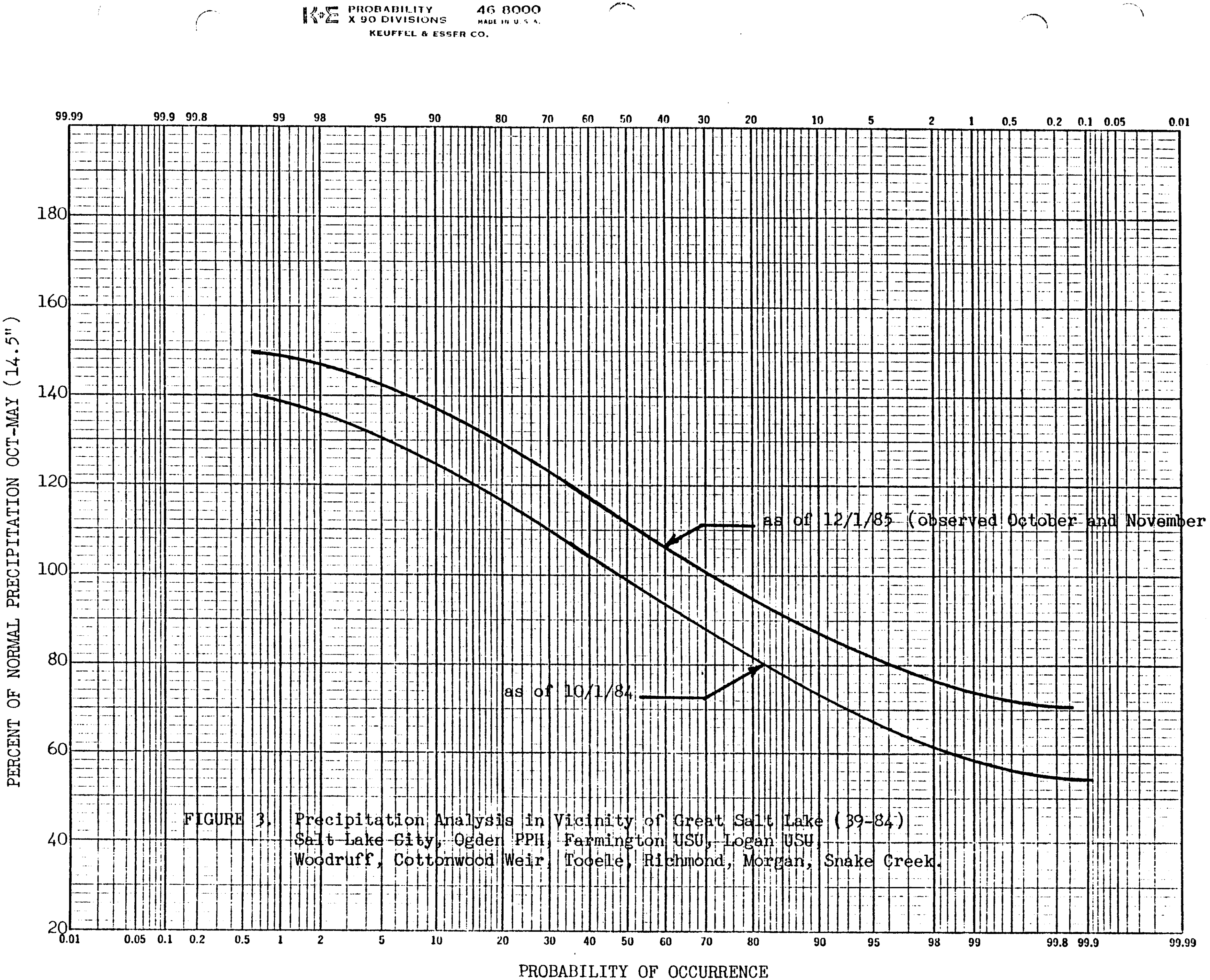

FIGURE 3. Precipitation Analysis in Vicinity of Great Salt Lake (39-84)
Salt Lake City, Ogden PPH, Farmington USU, Logan USU,
Woodruff, Cottonwood Weir, Tooele, Richmond, Morgan, Snake Creek.

PROBABILITY OF PRECIPITATION AND LAKE LEVEL 1985.

To maintain the current 1985 lake levels using an average depth of evaporation requires a dramatic change to a significantly wetter climatic regime. Again using information shown in Figure 2, the relationship between required long-term average precipitation at the Salt Lake City Weather Service Forecast Office and corresponding long-term lake levels is evident. Hydrologic modelling studies indicate that 125% of normal precipitation at Salt Lake City would be required each year to maintain an equilibrium lake level of 4210 feet MSL. This amount of precipitation would produce an equilibrium inflow of about 160% of normal. The graph presented by State of Utah Water Resources Division shows a stable lake level of 4209 feet MSL with an equilibrium water amount of 150% of average inflow (Figure 4), and using Figure 2 we find that this would require about 19 inches of precipitation per year, which is 115-125% of normal precipitation.

Other similar graphs by Eckoff, GSLM and the State of Utah State Department of Water Resource show a somewhat consistant relationship.

From the precipitation probability relationship in Figure 3 and the graph shown in Figure 2, the low probability of maintaining high lake levels can be estimated. It is evident that precipitation amounts that have occurred relatively infrequently in the past would have to occur year after year to maintain the lake at current levels. The probability of consecutive rare events occurring is not shown here but it is emphasized that a dramatic reduction in likelihood occurs compared to a single rare event.

Figure 5 shows the relationship between the percentage of precipitation and the expected peak lake level only for 1985. This is calculated from modelling which uses antecedent conditions, current inflow, current snowpack, current lake temperature and salinity, etc. It is advantageous to combine relationships such as shown in Figure 3 and 5 into the relationship shown in Figure 6. This relationship shows the probability of occurrence in 1985 for a given maximum lake level. (It is possible to obtain approximate 1986 probabilities from this graph by using a contingency of normal precipitation through May 1986 and an assumed lake level and an assumed antecedent condition on October 1, 1985.)

The precipitation occurring subsequent to a forecast date is the major unknown parameter with the most variability. Evaporation, rate of snowmelt, and the nature of precipitation can also significantly effect the seasonal rate of change in lake level. The relationships in the subsequent months among evaporation, cloud cover, relative humidity, temperature, precipitation, water temperature, etc., are impossible to project. However, precipitation is the dominant factor determining lake level and somewhat reliable frequencies of occurrence can be made. Therefore the integration of all parameters into a hydrologic water budget leads to a physically sound and understandable analysis. The most desirable feature of this method seems to be the ability to use measurable parameters and to apply solid physical relationships.

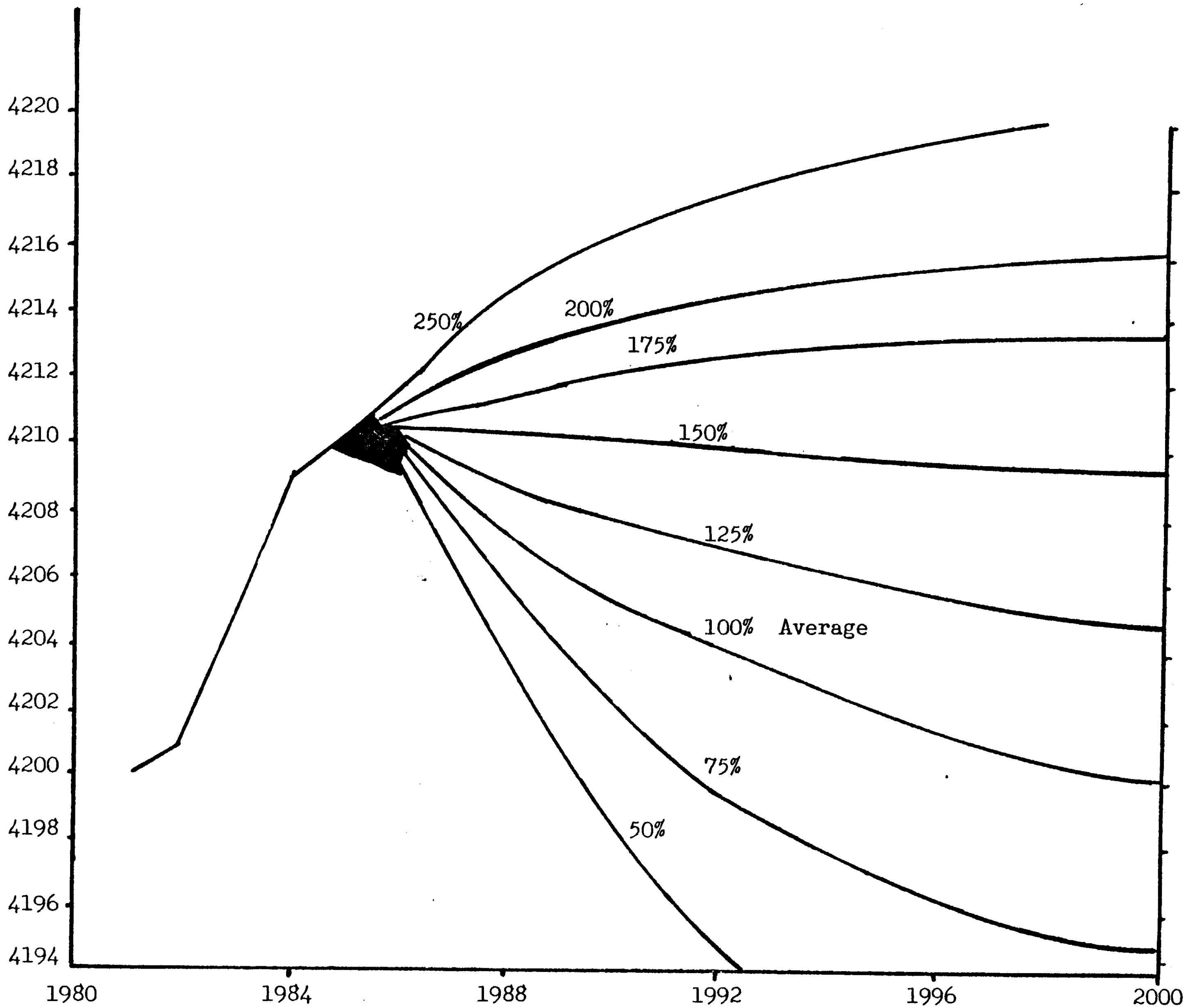

FIGURE 4. EquilibriumLevels of Great Salt Lake given amounts of Inflow: from Norm Stauffer, Utah State Department of Water Resources. 1984.

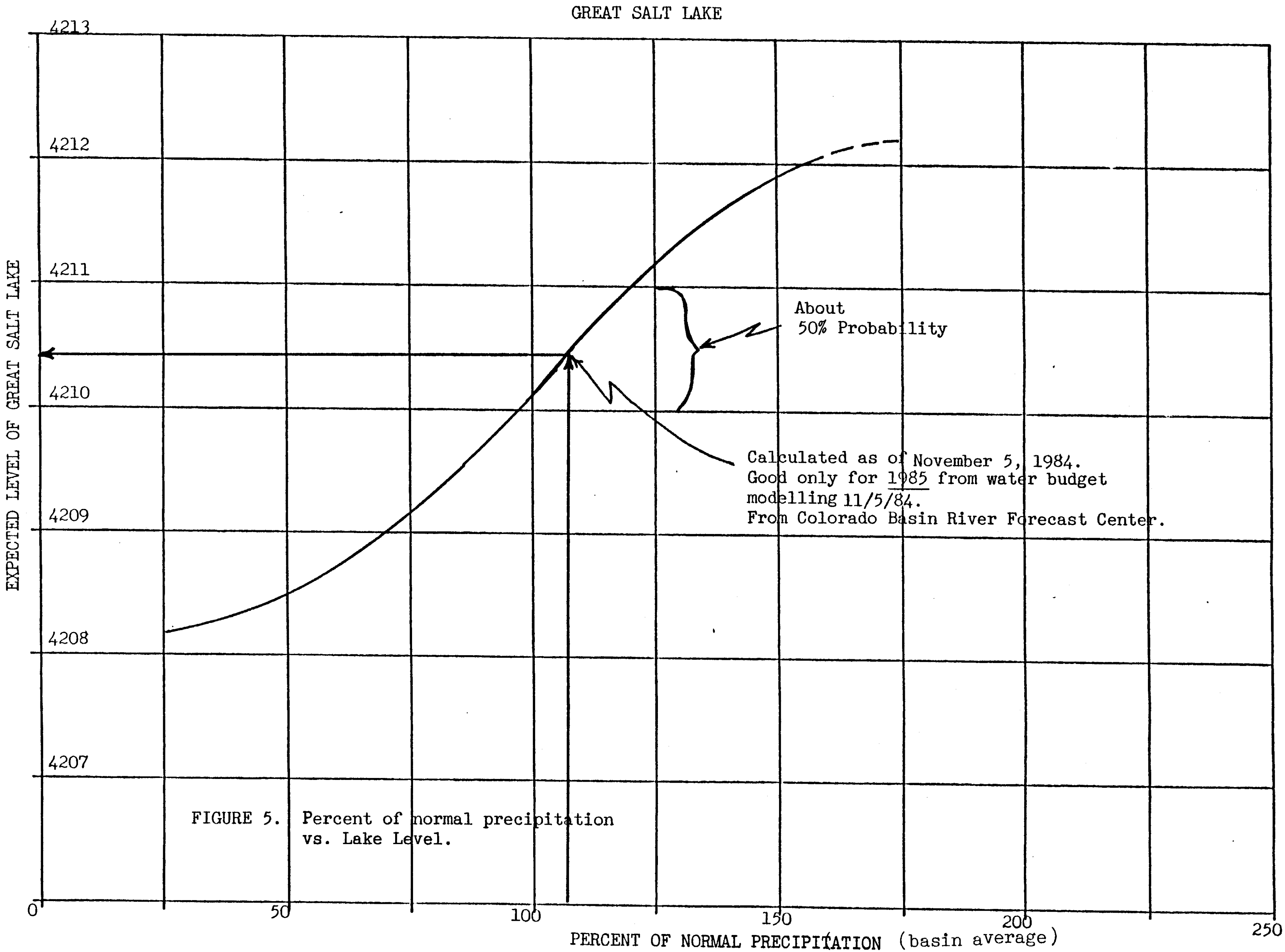

FIGURE 5. Percent of normal precipitation vs. Lake Level.

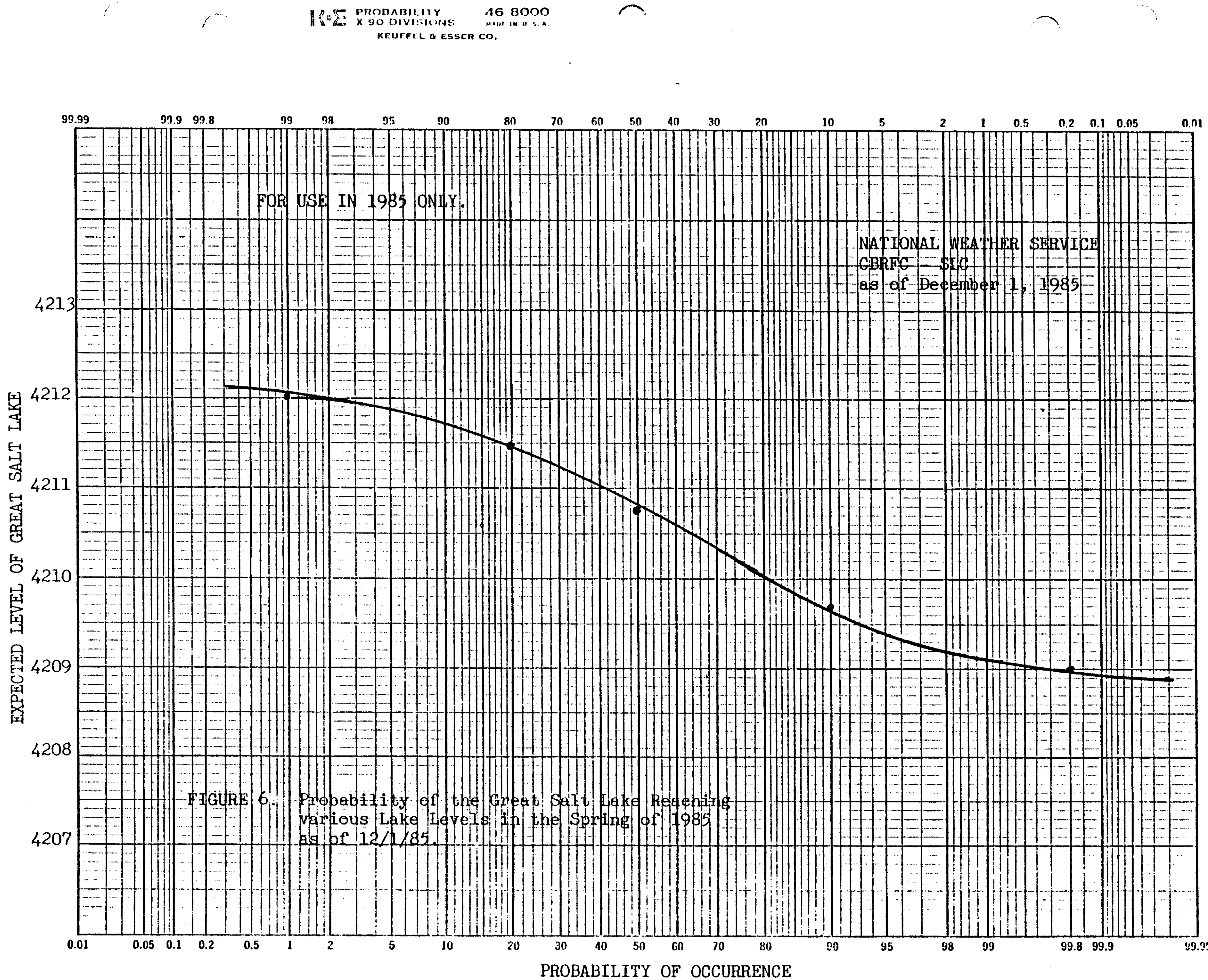

FIGURE 6. Probability of the Great Salt Lake Reaching various Lake Levels in the Spring of 1985 as of 12/1/85.

Contingencies using various precipitation amounts into the future allows realistic estimates of various lake levels in the next 1-2 years and the probability of achieving each projected lake level. Any model that attempts to predict future lake levels may be invalid if it does not use precipitation and subsequent runoff as the primary variables affecting lake level.

EVAPORATION

Evaporation, which relates directly to lake losses as precipitation does to lake inflow, has a dramatic impact on the levels. In dry years an assumption can be made that spring and summer are warmer with less humidity and less cloud cover and thus more evaporation. On the other hand in wet years, more cloud cover is noted in the spring and summer and humidity is increased, thus decreasing evaporation, as shown in Figure 7. In 1983 and 1984 evaporation may have been as much as 1/2-1 foot below normal on the Great Salt Lake. This may result in nearly 1/2 foot per year of additional lake rise due to reduction of evaporation. Figures 8 and 9 show some relationships between precipitation, evaporation, and volume changes in the lake at various elevations. These are presented to show the dynamic nature of the lake evaporation. The lake tends to be self regulatory. As the area increases more inflow is required to maintain the increased level and as the lake reaches a low point less inflow is required to maintain a given level. At the current lake levels, evaporation is near 5 million acre-feet (Figure 9) per year--this exceeds the long-term average inflow by 1-2 million acre-feet. To maintain the lake at 4210. feet MSL, average inflow plus the additional 1-2 million acre-feet is required. The graphs presented here indicate the changes in volume of losses due to evaporation and changes of inflows due to direct precipitation on the lake. The conclusions are apparent and dramatic. At the low point in 1963 of 4191.35 feet, total inflow due to average precipitation on the lake would be less than 600,000 acre-feet and evaporation loss would be near 2 million acre-feet (assuming 45" evaporation depth per year), whereas in 1985 at an elevation of 4210.0 feet, average precipitation on the lake would produce an inflow in excess of 1.6 million acre-feet and evaporation (45") would produce a volume loss of near 5.5 million acre-feet.

The evaporation losses adjusted for salinity are also plotted in Figure 9. These are losses assuming atmospheric conditions remaining constant and yielding 100% of potential evaporation from fresh water and 75% of potential from water saturated with salt.

IN SUMMARY

The Great Salt Lake has varied dramatically in size of surface area during the last 150 years, ranging from a low near 600,000 acres to a high near 1 1/2 million acres. Several factors affect variations in lake volume from year to year, and combine to complicate the analysis of the variance of the lake levels. Therefore it is important to remember that lake level readings (stages) are not as good an indicator of climatological variations as the study of total volume changes in the lake.

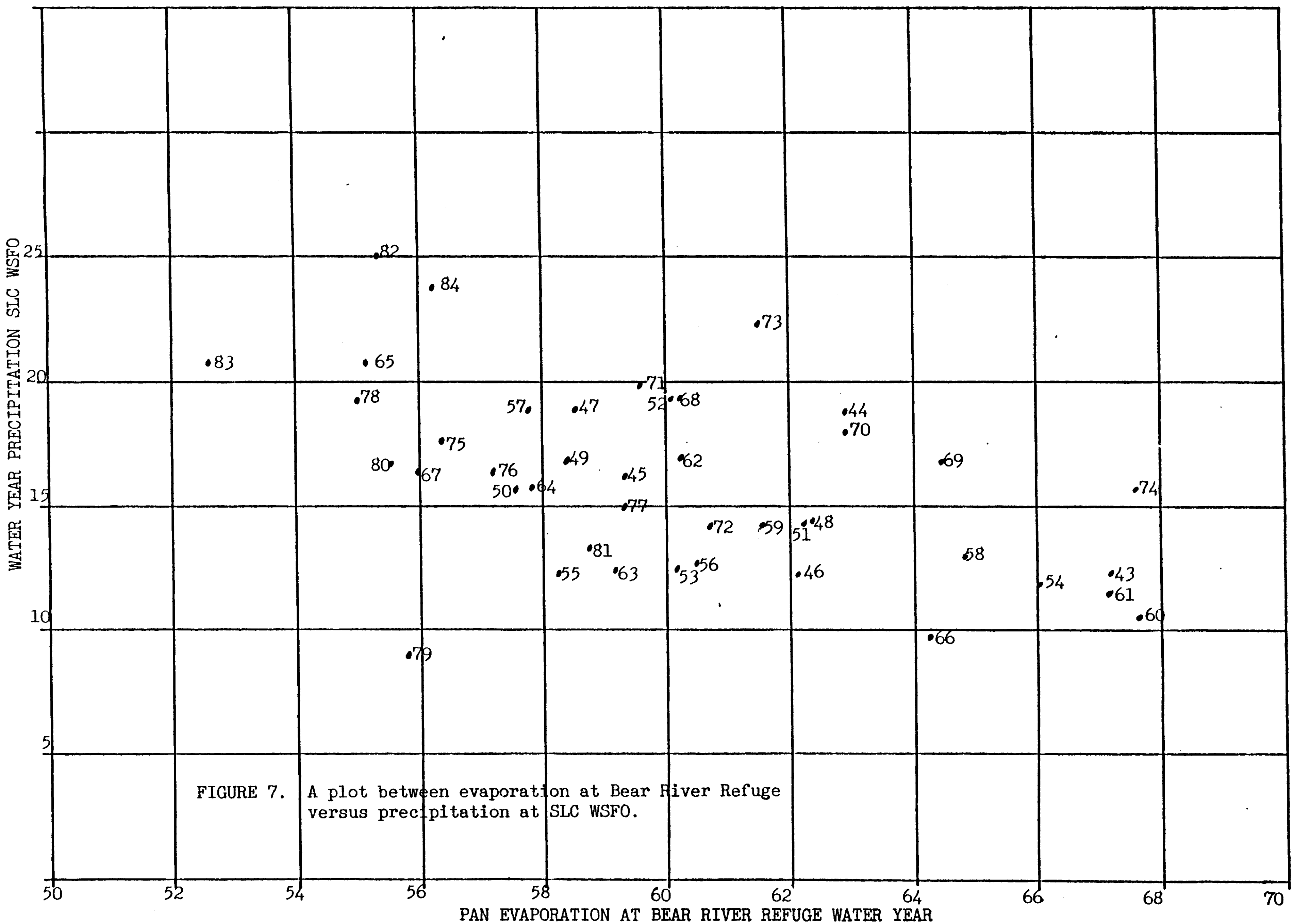

FIGURE 7. A plot between evaporation at Bear River Refuge versus precipitation at SLC WSFO.

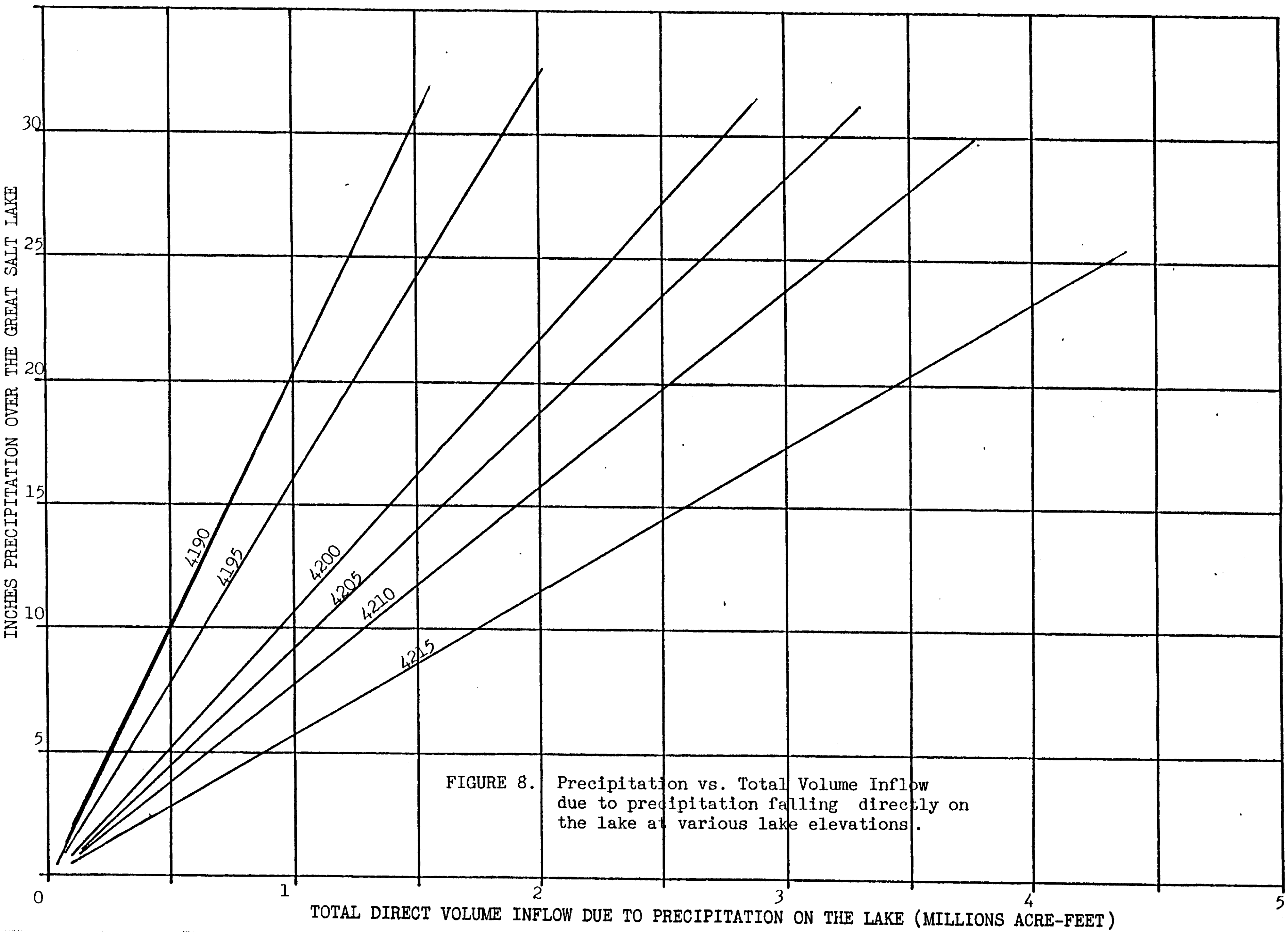

FIGURE 8. Precipitation vs. Total Volume Inflow due to precipitation falling directly on the lake at various lake elevations.

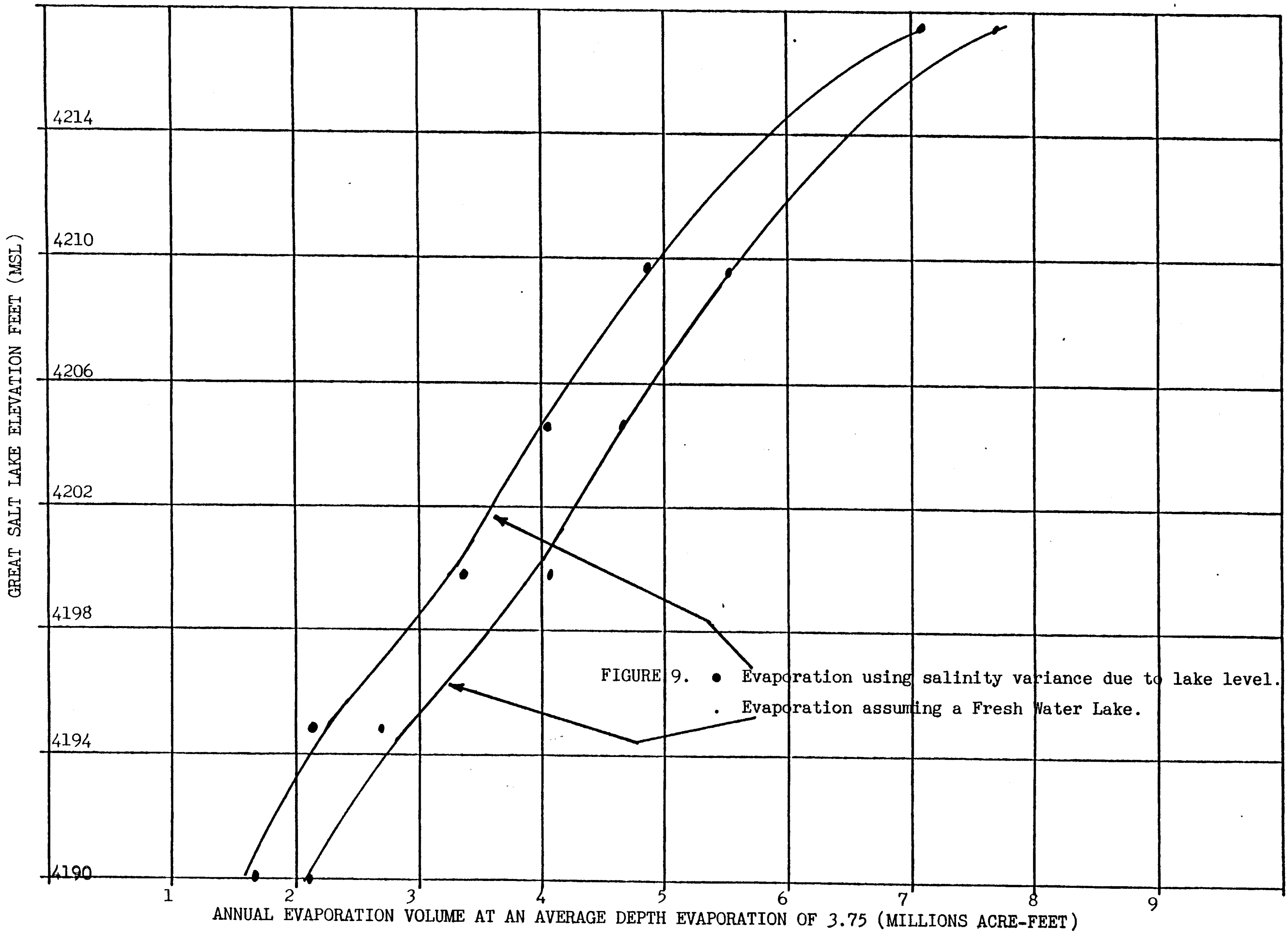

FIGURE 9. ● Evaporation using salinity variance due to lake level.
. Evaporation assuming a Fresh Water Lake.

Information has been presented relating volume changes to lake elevation and the physical relationships between sound hydrologic paramaters. These have shown that past lake levels are an excellent method to determine past climate but physically sound hydrologic parameters seem to be the best tool for estimating lake levels in the near future.

The rate changes since 1982 have been dramatic and unprecedented in the last 150 years. To maintain such an anomolously high inflow volume each year would require a drastic change in the climate around the Great Salt Lake. Average precipitation over the last 105 years (15.71 inches) has produced an average lake level of 4200.63 feet MSL. To maintain the level of 4210.5 feet MSL would require 125% of the long-term average precipitation to occur each year. This is extremely improbable. However, today's saturated conditions will allow slightly less precipitation (i.e., less than 125% of average) to maintain this level for the next few years. For the lake to increase in size would require precipitation in excess of 125% of average for an extended period.

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SHORT TERM LAKE LEVEL PREDICTIONS FOR THE GREAT SALT LAKE

by

E. Eric Adams, Rafael L. Bras, Donald R. F. Harleman,
Carlos Puente and Claire M. Bischoff

Department of Civil Engineering, Massachusetts Institute of Technology
Cambridge, MA 02139

INTRODUCTION

During the 1983 water year, the Great Salt Lake rose nearly five feet to reach a peak elevation of 4204.7 feet above MSL, the highest level in the previous 60 years. During water year 1984 the level rose an additional four and one-half feet to 4209.3, the highest level in over 100 years. The first three months of water year 1985 have again been wetter than average suggesting the possibility of a third straight year of record water levels.

The purpose of this work is to predict water levels over the next six years using a lake water balance. The approach differs from previous studies in two major aspects. First, evaporation is predicted by combining dynamic energy and salt balance models with the water budget model, allowing interactions among evaporation, temperature and salinity, and thus reducing errors in predicted evaporation. Second, future annual inflows are predicted using a uni-variate Broken Line Model, which models observed long-term persistence in the historical streamflow data through specification of the Hurst coefficient. Monthly values of hydrological and meteorological variables are derived from annual streamflow through a multi-variate disaggregation model. Prediction of future lake levels is based on simulations using 50 sequences of synthetic input. Additional details are found in Adams, et al. (1985).

DYNAMIC WATER-ENERGY-SALINITY BUDGET ANALYSIS

Equations The model consists of coupled equations expressing conservation of mass, thermal energy and salt. Each balance is applied separately to south and north basins, which are assumed to be well-mixed with respect to temperature and salinity.

Conservation of mass is expressed

$$\frac{d}{dt}(\rho_s \forall_s) = \rho_f Q_i + \rho_f P_s A_s - \rho_f E_s A_s + \rho_n Q_2 - \rho_s Q_1 \qquad (1a)$$

$$\frac{d}{dt}(\rho_n \forall_n) = \rho_f P_n A_n - \rho_f E_n A_n + \rho_s Q_1 - \rho_n Q_2 \qquad (1b)$$

where t = time; ρ = density; $\forall$ = volume; A = surface area; Q_i = freshwater inflow rate (comprising Jordan, Weber and Bear rivers); E = evaporation rate; P = precipitation rate; Q_1, Q_2 = interbasin exchange flows from south to north and north to south; and subscripts s, n, and f refer to south basin, north basin, and fresh water.

Conservation of thermal energy is written

$$\frac{d}{dt}(\rho_s c_s \forall_s T_s) = \rho_f c_f Q_i T_i + \rho_f c_f P_s A_s T_p - \rho_f c_f E_s A_s T_s$$
$$+ \rho_n c_n Q_2 T_n - \rho_s c_s Q_1 T_s + \phi_{net,s} A_s \qquad (2a)$$

$$\frac{d}{dt}(\rho_n c_n \forall_n T_n) = \rho_f c_f P_n A_n T_p - \rho_f c_f E_n A_n T_n + \rho_s c_s Q_1 T_s$$
$$- \rho_n c_n Q_2 T_n + \phi_{net,n} A_n \qquad (2b)$$

where T = basin-average temperature; c = specific heat; supscript p refers to precipitation; and ϕ_{net} = net surface heat flux consisting of five components,

$$\phi_{net} = \phi_{sn} + \phi_{an} - \phi_b - \phi_e - \phi_c \qquad (3)$$

where ϕ_{sn} = net solar (short wave) radiation, ϕ_{an} = net atmospheric (long wave) radiation, ϕ_b = back radiation, ϕ_e = latent heat flux due to evaporation and ϕ_c = sensible heat flux.

Established formulae are used to compute each component from known lake and atmospheric properties. Separate calculations are made for each basin. ϕ_{sn} is computed using graphs published by Hamon et al. (1954) for daily average clear sky radiation as a function of latitude and time of year, and correcting for reflectivity (assumed 6%) and cloud cover (Wunderlich, 1972). ϕ_{an} is computed using Swinbank's (1963) clear sky formula, 3% reflectivity and a cloud cover correction (Geiger, 1965). ϕ_b assumes an emissivity of 0.97. ϕ_e and ϕ_c are computed using Harbeck's (1962) fetch-dependent bulk aerodynamic formula and the concept of a Bowen ratio relating sensible to latent heat flux. Vapor pressure is corrected for salinity using data of Dickson et al. (1965) for GSL brines. Evaporation for each basin is computed from the respective latent heat fluxes.

The last balance expresses salt conservation, which is written

$$\frac{d}{dt}(\rho_s S_s \forall_s) = \rho_n S_n Q_2 - \rho_s S_s Q_1 \qquad (4a)$$

$$\frac{d}{dt}(\rho_n S_n \forall_n) = \rho_s S_s Q_1 - \rho_n S_n Q_2 \qquad (4b)$$

where S = salinity. The exchange flows Q_1 and Q_2 between basins consist of free surface flow through the culvert and flow through the porous media beneath and within the causeway. As a first step, net flow from south to north is computed in the form

$$Q_{net} = C_e (\Delta h - \Delta h_e)^{1/2} \qquad (5)$$

where $\Delta h = h_s - h_n$ = the elevation difference between south and north; and Δh_e = the equilibrium elevation difference where no net flow would occur. Using estimated exchange flows for 1971-1972 (Waddell and Bolke, 1973) and observed basin level differences, Δh_e = 0.75 feet and C_e = 2000 $ft^{2.5}/s$ were estimated for conditions prior to August 1984. After August 1984, when the causeway was breached, a value of C_e = 6000 $ft^{2.5}/s$ was used. Q_1 and Q_2 were computed by assuming that the total salt content in each basin remains constant and requiring that Q_{net} = $Q_1 - Q_2$ leading to

$$Q_1 = \frac{Q_{net}\,\rho_n\,S_n}{\rho_n\,S_n - \rho_s\,S_s} \qquad\qquad Q_2 = \frac{Q_{net}\,\rho_s\,S_s}{\rho_n\,S_n - \rho_s\,S_s} \tag{6}$$

Solution Procedure Starting with known initial conditions, the above equations are solved using an explicit time scheme. Area and elevation are computed from updated volume by interpolation from curves of area and volume vs. elevation for each basin obtained from the Utah Geological and Mineral Survey (UGMS). These curves are valid up to elev. 4218 but have been extrapolated to elev. 4222 to simulate extreme flooding events. The model uses monthly data for all independent variables; however, a numerical time step of 0.2 months was selected to insure smooth results.

Data To simulate past events, the model requires monthly average data for the following variables: Q_i, T_i, P, T_p, T_a, ϕ, W_2, and C.

Freshwater inflow (Q_i) consists of the sum of the inflows from the Jordan, Weber and Bear rivers. Explicit contributions from direct run-off and groundwater are neglected. Data for the three rivers are based on USGS records near Salt Lake City (Jordan R.); near Plain City (Weber R.); and near Corinne (Bear R.). Inflow temperature (T_i) is taken from USGS records on the Weber River as the average of water years 1979-1981.

Precipitation (P) is taken as the weighted average of data from Ogden, Corinne, and Salt Lake City airport (SLC). T_p is assumed to equal air temperature. Air temperature (T_a), relative humidity (ϕ), wind speed (W_2) and cloud cover (C) are taken from SLC data summarized in Local Climatological Data publications of the National Weather Service. Wind speed was converted to 2-meter elevation over water by calibration against data collected by AMOCO.

Measurements of elevation, temperature and salinity were required for initial conditions and for comparison during model calibration and verification. Monthly elevations for each basin were available for water years 1940-84, while approximately quarterly values of temperature and salinity for each basin were obtained from UGMS for 1968-84. Monthly averaged temperatures, taken by UGMS near the southern shore of GSL from June 1977 through July 1978, provided additional data for verification.

Sensitivity Studies Complete data sets were available for the water years 1969-84 for purposes of model calibration and verification. In the sensitivity phase, the model was run from October 1968 through 1972 and results were compared with measurements from 1970 to 1972. Table 1 summarizes the performance of this run (designated Run 1; base case) in terms of the mean error and standard deviation for weighted average elevation and temperature. Averaging is based on 2/3 (south) plus 1/3 (north) and is performed to filter out any error in predicted elevation differences and corresponding exchange flows between basins.

Before calibration, a sensitivity study was performed to identify the model response to several input parameters, including a 10% increase in net incoming radiation (ϕ_{sn} and ϕ_{an}), a 20% increase in wind speed (W_2), and a 20% increase in total inflow (Q_i and P). Results are also tabulated in Table 1 and lead to the following general observations. Increasing radiation (Run 2) directly increases water temperature, which indirectly increases evaporation, leading to a decrease in elevation. Note that sensitivity to radiation includes sensitivity to T_a and C which are used to calculate ϕ_{sn} and ϕ_{an}. Increasing wind speed (Run 3) directly increases evaporation, which leads to decreased elevation and temperature. Note that sensitivity to wind speed is equivalent to sensitivity to the evaporation equation and also combines sensitivity to other independent variables (T_a and ϕ), as well as dependent variables (S). Finally, increasing inflow rate (Run 4) directly increases elevation but has a negligible effect on temperature. Note that this also tests the sensitivity to possible groundwater or ungauged surface water sources.

Initial calibration involved adjusting input parameters by a fixed fraction, following the sensitivity studies. Because three inputs were varied (radiation, wind speed and inflow) while only two outputs (elevation and temperature) were analysed, more than one set of adjustments was possible. The adjustments were made to reduce the anticipated mean errors of the two outputs to zero (based on the sensitivity studies) while simultaneously minimizing the sum of the squares of the percentage adjustments. The optimization resulted in a 5% decrease in radiation and no change in wind speed or inflow. Results of the calibrated model are shown as Run 5 in Table 1, and indicate an 86% reduction in absolute mean error for weighted average elevation and a 100% reduction in mean absolute error for weighted average temperature.

Verification and Final Calibration The model was re-run using the above calibration for the entire period 1968-84. Results are shown in Figures 1 and 2 which show computed (crosses) versus measured (circles) southern basin lake temperatures and weighted average annual peak elevations.

As with the calibration run, temperatures show reasonable mean agreement with a tendency to overpredict in the summer and underpredict in the winter. In comparison with the detailed temperature survey of 1977-78 (open circles in Figure 1), the mean error is only 0.3°F.

Elevation predictions show more discrepancy. While good overall agreement is seen over the first four years, predicted peak elevations begin to lag the measurements in 1973, and by 1983 and 1984, the discrepancy has grown to several feet. Clearly, the constant calibration factors identified previously are not sufficient.

Closer examination of Figure 2 suggests that the error in predicting year-to-year peak elevation is correlated with annual inflow. During years with large inflow (e.g., 1983, 1984), the model underpredicts the rise in water level, while during years with small inflow (e.g., 1970, 1977), the model underpredicts the magnitude of decrease. Because water temperature is generally well predicted, the error is most likely in total inflow. Accordingly, in Figure 3 "actual" annual inflow Q_a, as deduced from the model water balance, is plotted against measured annual inflow Q_i for the same year. Regression analysis shows that Q_a and Q_i (in units of million acre-ft/yr) are related according to

$$Q_a = -.59 + 1.41\ Q_i \qquad (7)$$

with r = .98. The fact that $Q_a > Q_i$ during wet years is attributed to ungauged surface run-off and to groundwater flow.

To further improve the correlation, the residual error ΔQ_a (difference in Q_a back computed with the water budget and with Eq. 7) was correlated with previous years' measured Q_i, lake elevation, and time. The best correlation was found with time. Including this correlation, the corrected annual inflow (in million acre-ft/year) is given by

$$Q_a = -1.00 + 1.41\ Q_i + .051\ (\text{Year}-1969) \qquad (8)$$

The model was re-run using Eq. 8 to adjust Q_i. The adjustment has only a small effect on predicted temperature. Results for peak elevation are shown as pluses in Figure 2 and indicate substantial improvement; after 15-16 years of prediction, the error in predicted peak elevation is less that 0.1 ft in 1983 and 0.4 ft in 1984. Comparison on a bi-monthly basis (not shown) indicates that the model also predicts realistic seasonal changes in elevation.

Evaporation Predictions Table 2 summarizes evaporation for each of the 16 years of simulation. Average annual evaporation in the northern basin (29.5 in/yr) is less than the rate in the southern basin (34.0 in/yr), as expected, due to increased salinity. "Freshwater" evaporation for the two basins is presented in columns 3 and 4. Ratios of annual mean saltwater to "freshwater" evaporation rates are 80% and 67% for the south and north basins, respectively; ratios for the last two years are somewhat higher than average, reflecting decreasing salinity due to higher water levels. "Freshwater" is used in quotes because these rates are based on computed water temperatures which reflect computed salinities. Because evaporation decreases with salinity, computed water temperatures (as well as computed "freshwater" evaporation rates) are greater than in an equivalent freshwater lake. (This also explains why the "freshwater" evaporation rate is higher in the north than in the south.)

Computed evaporation cannot be compared with direct measurement of lake evaporation, but a comparison can be made against other estimates. NOAA's Climatic Atlas (U.S. Dept. of Commerce, 1977) shows mean annual lake evaporation of 42 in/yr based on data from 1946-1955. This is slightly lower than our computed averages of 42.7 and 43.9 in/yr based on 1969-84; this is consistent with the fact that Climatic Atlas estimates are based on freshwater temperature, whereas our values reflect saltwater temperature.

Finally, the last column of Table 2 lists computed freshwater evaporation for 1969-77 taken from Table 15 of James et al. (1979). These values are based on a long term average freshwater evaporation rate of 52 in/yr (determined by a Utah Div. of Water Resources water balance), and adjusted for annual variation using freshwater pan data from Corinne. James et al. (1979) determine saltwater evaporation rate by multiplying the freshwater rates (column 5 of Table 2) by a ratio of saltwater-to-freshwater evaporation as a function of salinity. Predictions in columns 3-5 of Table 2 show similar trends with time. However, our computed "freshwater" evaporation rates are about 15% lower than those of James et al. (1979) and show less year-to-year variability. Our smaller variability reflects coupling between temperature and evaporation; e.g., during periods of high evaporation, water temperatures drop, reducing evaporation, and vice versa.

STOCHASTIC INPUT GENERATOR

Data included monthly values of streamflow of major rivers going into the lake (1943-84), precipitation at several locations (1943-84), snowpack at several stations tributary to the rivers of interest (1943-84), and cloud cover, air temperature, relative humidity, and wind speed obtained from SLC climatological records (1969-84).

Extensive statistical analysis was performed at annual and monthly levels but space limitation allows only a brief summary. Table 3 gives the annual mean, standard deviation, skewness, lag-one autocorrelation coefficient, and Hurst coefficient for most variables. Snowpack at different locations and times of the year was compared with annual and monthly streamflows to identify the best correlations.

All variables exhibited significant seasonality. The concurrent cross-correlations among total streamflow, precipitation, humidity, wind speed, cloud cover, and temperature for different months were significant.

When transformed by subtracting monthly means and dividing by standard deviations, all variables, except streamflow, showed little autocorrelation from month to month as well as insignificant cross correlations with other variables for time offsets greater than zero. The correlations between total annual streamflows and streamflows at different months of the year as well as the correlation among months in a year were significant.

Based on the statistical analysis, the following guidelines to model building were established: 1) Long-term persistence of the annual total streamflow is important and should be preserved in any multi-year prediction; 2) Monthly streamflows exhibit significant correlations that must be maintained in any simulation; 3) Monthly precipitation is best obtained via regression on concurrent monthly streamflow; 4) Wind speed, humidity, cloud cover, and temperature show no significant time structure beyond the dominant seasonality. Once this is accounted for, the residuals are random variables, some significantly correlated to others and with precipitation. This behavior should be preserved.

Simulation Algorithm The above guidelines led to the following algorithm.

1. Annual total streamflow was simulated for six years into the future. Simulations were conditioned on observed streamflow for each of the past two water years. 50 different scenarios of six-year data were generated.
2. The total annual streamflows, for all scenarios, were then disaggregated into monthly values. Hence 50 scenarios of 72 months were obtained.
3. Concurrent monthly precipitation, wind speed, cloud cover, temperature and relative humidity were obtained preserving cross-correlation among them and the monthly streamflows.
4. For predictions beginning October 1, 1984 (beginning of water year), the resulting 50 scenarios of monthly values for the next 72 months were input into the water-energy-salinity budget model resulting in a distribution of peak lake levels in each of the next six years.
5. For predictions made on January 1, 1985, snowpack, rainfall, and streamflow through January 1st were used to obtain, via regression, a new estimate of the remaining total streamflow in the next nine months. This lead to a new estimate of the annual total streamflow which was combined with the October 1st prediction to obtain a refined estimate. This new estimate was disaggregated into monthly values. Steps 3 through 4 were then repeated for the coming nine months, leading to an updated prediction of the current water year peak lake level.
6. Similar prediction can be made on April 1, 1985. Step 5 would be repeated with observed snowpack and streamflow through March 31, 1985, leading to a final peak lake level elevation prediction for the year. To make predictions on October 1, 1985, the whole algorithm, starting at Step 1, would be repeated.

Chosen Models Clearly the prediction algorithmn is "driven" by the simulated streamflow; hence the choice of model will significantly influence everything else. Two desired characteristics dominated model choice: preservation of long-term persistence, and ability to condition on past streamflows. Obviously the model must also perform well in preserving other important statistics.

Long-term persistence refers to low frequency behavior such that streamflows remain higher or lower than mean values for periods (and magnitudes) not expected from traditional time series models. The parameter commonly used to describe this behavior is the Hurst coefficient. As Table 3 shows, the Hurst coefficient, H, of the total inflow is considerably above 0.5, which could be caused by a non-stationary mean, which usually is difficult to infer from data, or a stationary process with unusually long memory. Pragmatically, the latter hypothesis is usually assumed in modelling Hurst behavior. Bras and Rodriguez-Iturbe (1984) discuss the philosophy and the nature of existing models.

In order to preserve the observed Hurst coefficient several alternatives, including autoregressive, ARMA(1,1) (O'Connell, 1974) and autoregressive Markov (Lettenmaier and Burges, 1977) models were investigated. None satisfactorily preserved the combination of mean, variance, skewness, Hurst coefficient and lag-one correlation. Finally the well established Broken Line Model was selected and implemented using the computer code developed by Curry and Bras (1978). Table 4 compares the statistics computed by the model in comparison with measured statistics.

Unfortunately the structure of the Broken Line Model does not allow for simple, explicit, conditioning on past values. This objective was achieved by simulating 500 series each of length 42 years (to match the length of the historical sequence), and searching the record for eight-year segments that began with sequential values resembling 1983 and 1984 streamflows. Resemblance was established if the simulated flows were within a percentage of the actual flows. The smaller the percentage, the smaller the number of resembling segments which could be found.

Table 5 gives the conditional histogram, plus mean and standard deviations, of the flows of years 1985-90 when the resemblance criteria was 15%. Table 5 is based on 50 identified segments starting with a pair resembling the 1983-84 sequence. As expected, flows conditioned on the high 1983 and 1984 occurrences show mean values considerably higher than historical variances -- even 5 years into the future. However, by 1990 the means approach those of the historical data (Table 3).

Annual discharges were disaggregated into monthly flows following Valencia and Schaake (1972),

$$y = \bar{y} + a(x - \bar{x}) + bw \qquad (9)$$

where y is the vector of monthly flows with mean $\bar{y}$, and x is the corresponding annual flow with mean $\bar{x}$. Parameters a (12x1 vector) and b (12x12 matrix) allow monthly flows to preserve cross correlations among themselves and the yearly flows. The term w represents a random component with zero mean and unit variance.

Monthly values of P, T_a, W_2, C and ψ were obtained by using a monthly regression that relates them to monthly discharge. The form of the regression model is similar to that of the disaggregation model, i.e.,

$$y_i = \bar{y}_i + a_i(x_i - \bar{x}_i) + b_i w_i \qquad (10)$$

where y_i is the vector of unknown meteorological values for month i with mean $\bar{y}_i$, and x_i are the corresponding discharges for month i with mean $\bar{x}_i$. The model parameters a_i (5x1 vector) and b_i (5x5 matrix) are obtained so that cross correlation among the variables of interest is preserved. A log transformation of precipitation values was made to insure generation of positive values. Generated meteorological values were obtained by introducing the generated monthly discharge values into Eq. 10 and generating the error terms from a standard Gaussian variate.

Prediction Update As information on January 1, 1985, and April 1, 1985, became (becomes) available, an improved estimate of the total annual flow for the current year could (can) be obtained. A model that produces such an estimate is:

$$\hat{X} = \overline{X} + \sum_{i-1}^{m} a_i(Z_i - \overline{Z}_i) + bw \tag{11}$$

where $\hat{X}$ is the estimator of total annual flows, $\overline{X}$ is the mean annual flow, Z_i are observed values of the current year of, e.g., discharge (cumulative over three or six months), precipitation and snowpack with respective long term means $\overline{Z}_i$. Parameters a_i and b allow the cross correlations among the explanatory variables and total annual discharge to be preserved. By assuming w to be standard Gaussian, the estimator X also becomes Gaussian with mean $\overline{X} + \sum a_i(Z_i - \overline{Z}_i)$ and variance b^2.

An improved estimate of the distribution of annual flows is obtained by combining the distribution of the above estimator with that of the generated values based on the previous October 1st. This was done by weighting the distributions according to their respective variances, i.e., by giving more weight to the estimate with less variance. The final estimate of the annual flow distribution, therefore, becomes,

$$f_X(x) = c_1 f_{\tilde{X}}(x) + c_2 f_{\hat{X}}(x) \tag{12}$$

where $f_{\tilde{X}}(x)$ is the distribution of generated annual flows using the Broken Line Model and $f_{\hat{X}}(\hat{X})$ is the distribution generation by Eq. 11. c_1 and c_2 are given by

$$c_1 = \frac{\sigma_{\hat{X}}^2}{\sigma_{\hat{X}}^2 + \sigma_{\tilde{X}}^2} \qquad c_2 = \frac{\sigma_{\tilde{X}}^2}{\sigma_{\hat{X}}^2 + \sigma_{\tilde{X}}^2} \tag{13}$$

Representative time series of discharge for the remaining nine or six months are obtained by sampling the above distribution, (Eq. 12), and subtracting from them the observed cumulative monthly flow. Updated monthly flows for the remainder of the year are obtained again by using the disaggregation model of Eq. 9. Once monthly flows are obtained, the remaining meteorological variables are found by Eq. 10.

Table 6 shows historical means as well as observed 1985 values of the explanatory variables employed in the regression for annual total streamflow. For comparison, observed values for 1984 are also included. Note that both years are considerably above the mean, but 1985 does not appear to be as extreme as 1984.

The regression formula for annual streamflow based on cross correlations among explanatory variables and total annual flows gives for 1985,

$$\hat{X} = 3.75 \times 10^6 + 0.77 \times 10^6 w \tag{14}$$

Table 7 gives a histogram with mean and standard durations of updated 1985 inflow.

PREDICTION OF FUTURE LAKE LEVELS

Calculations Starting October 1, 1984 Calculations began with conditions on October 1, 1984, when water levels were approximately 4206.8 (north) and 4207.8 (south). 50 simulations were made, each using 72 months of synthetic monthly input. The distributions of weighted average (2/3 south plus 1/3 north) peak elevations are summarized in Table 8. The median (50%) and 10% exceedance predictions are also summarized. A similar summary (not shown) indicated that the median prediction for the southern basin is .3 to .4 foot above the median weighted average prediction. In both cases, median predictions peak in the year 1987 or 1988 (at elevation 4213.9 for weighted average and 4214.3 for the south) while 10% exceedance predictions are still rising by 1990.

These calculations reflect the calibration of inflow rates according to Eq. 8, which includes a time-dependent term. In the calibration this regression equation provided the best fit. However, it is clear that the same time-dependence cannot persist indefinitely. To examine sensitivity, the 50 simulations were re-run with the time-independent inflow calibration of Eq. 7. Predicted peak elevations were lower than corresponding predictions with the time-dependent calibration by approximately 0.2' per year starting in 1985. For example, whereas the median predicted weighted average peak in 1988 was 4213.9' with the time-dependent calibration, it was 4213.1' with the time-independent calibration. For later years, the correct calibration is likely to fall between these two cases, but for the near term, the time-dependent calibration is preferred.

Calculations Starting January 1, 1985 Calculations for 1985 were updated using initial conditions on Jan 1, 1985, when water levels were approximately 4207.6 in the north and 4208.6 in the south. 50 simulations were made, each using 69 months of synthetic monthly data. The distribution of weighted average peak elevations, plus median (50%) and 10% exceedance estimates are summarized in the last column of Table 8.

Comparison with the results of the second column suggests that the updated peak elevations for 1985 are lower and less variable than those begun on Oct. 1, 1984. The difference in median peak elevation is approximately .8 ft. and is attributable to two factors: 1) the observed water level on Jan. 1 was less than the median water level predicted for Jan. 1 based on Oct. 1 initial conditions, and 2) while the combined inflow observed for Oct.-Dec. 1984 was higher than average (about 1.9 times the historical average for these months), the mean of the stochastically generated inflows for these months was about 2.2 times the historical average. Thus predictions starting Oct. 1, 1984 assumed wetter conditions (on average) than those which have been shown to date, and which are reflected in the input used for the updated predictions.

The fact that, for 1985, updated predictions of elevation are lower than the original predictions suggests that the same trend would be true for later years. Thus the estimates for 1986-90 in Table 8 would appear to be upper bounds. Revised estimates for the later years could be made by updating the stochastic prediction of annual total streamflow by considering the current predictions for 1985 streamflow.

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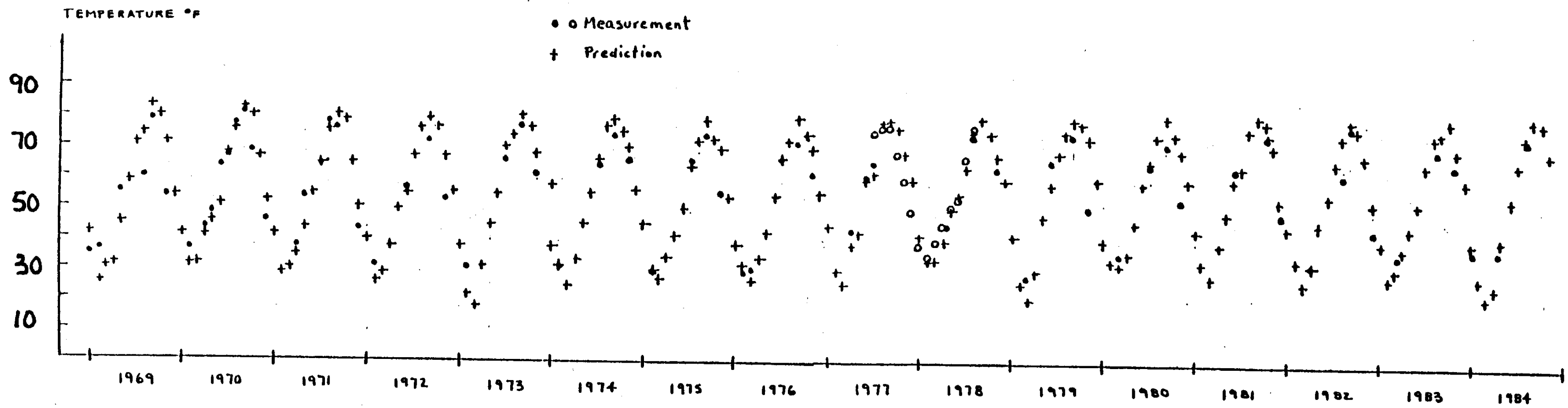

Figure 1 Measured versus Computed Lake Temperature (South Basin)

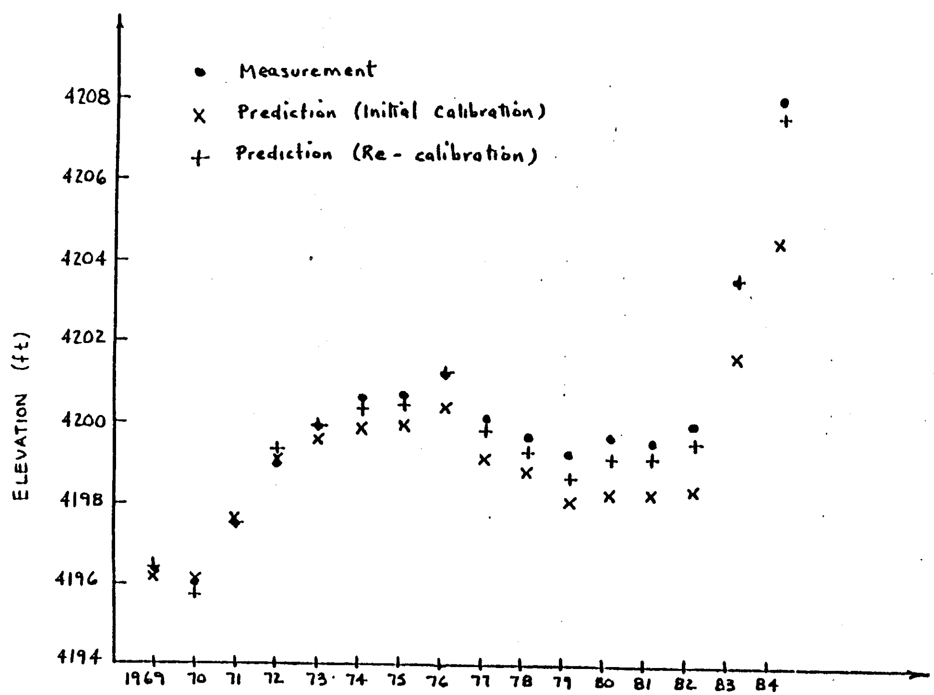

Figure 2 Measured versus Computed Annual Peak Weighted Average Elevations

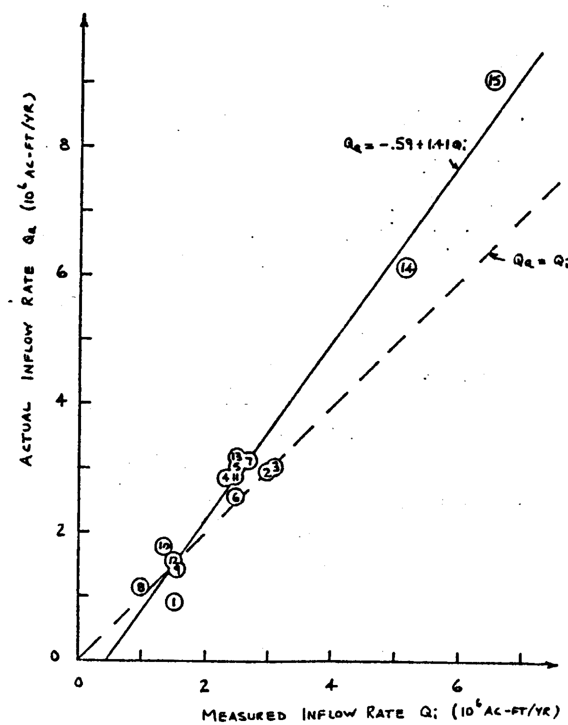

Figure 3 Correlation Between Actual and Measured Inflow Rate. Numbers refer to year after 1969 (e.g., 15 is 1984)

Table 1. Summary of Sensitivity, Initial Calibration and Verification Runs.

Run	Fractional change from base case for: $\phi_{sn} + \phi_{an}$	W_2	$Q_i + P$	Error (pred-meas) in Elevation (ft): mean	std. dev.	Error (pred-meas) in Temperature (°F): mean	std. dev.
1* (base case) 1970-72	0	0	0	-0.7	0.2	2.2	7.1
2* (sens) 1970-72	+.10	0	0	-2.2	0.5	7.0	7.3
3* (sens) 1970-72	0	+.20	0	-1.1	0.3	0.7	6.9
4* (sens) 1970-72	0	0	+.20	0.7	0.2	2.0	7.1
5* (init calib) 1970-72	-.05	0	0	0.1	0.1	0.0	7.1
6† (verif) 1977-78	-.05	0	0			0.3	6.5

*Sensitivity and Initial Calibration Runs start October 1968 and are compared with 13 data points for weighted average temperature and 18 (bi-monthly) points for weighted average elevation during 1970-1972.

†Verification Run starts October 1968 and is compared with 14 monthly data points for southern basin temperature during 1977-1978.

Table 2. Summary of Computed Annual Evaporation for 1969-1984.

Year	Pred. Evap. (in): South	North	Pred. "fresh-water" Evap. (in): South	North	James et al (1979)
1969	32.5	27.4	43.6	44.9	55.3
1970	30.0	25.6	41.3	42.6	53.0
1971	32.1	26.9	41.9	43.4	49.1
1972	35.7	30.4	44.5	45.9	51.0
1973	35.0	30.6	43.3	44.5	52.5
1974	37.8	33.1	45.8	46.8	58.3
1975	33.5	28.1	41.3	42.5	46.5
1976	35.4	31.2	42.6	43.8	49.6
1977	34.6	30.6	43.3	44.4	48.7
1978	34.3	30.1	43.5	44.7	
1979	35.4	31.4	45.6	46.5	
1980	33.9	29.2	42.9	44.1	
1981	33.5	29.6	42.8	44.2	
1982	32.7	28.2	40.9	42.4	
1983	34.3	29.2	40.3	41.9	
1984	34.6	30.8	39.5	40.2	
Ave.	34.0	29.5	42.7	43.9	51.6
St. Dev.	1.7	1.9	1.7	1.7	3.7

Table 3. Annual Statistics of Most Important Inputs

Variable	Mean	Standard Deviation	Skewness	Lag-1 Auto-Correlation	Hurst Coefficient
Total discharge (Jordan and Weber and Bear, acf)	1964004	1048782	2.30	0.544	0.834
Lake Precipitation (inches)	11.26	3.46	2.75	0.26	0.606
Temperature* (°F)	52.1	0.97	0.64	0.11	
Relative Humidity*	54.6	2.5	0.6	0.34	
Wind Speed* (mph)	9.1	0.3	-0.32	-0.20	
Cloud Cover* (0-10)	5.8	0.26	0.38	0.24	

*Annual Monthly Averages

Table 4. Statistics of Observed and Generated Streamflows

Flows	Mean	Standard Deviation	Skewness	Lag-one Correlation	Hurst Coefficient
Observed	1.96×10^6	1.0×10^6	2.3	0.72	0.834
Generated*	1.96×10^6	0.9×10^6	1.7	0.63	0.785

*Mean over 500 realizations each of 42 years

Table 5 Histograms of Generated Streamflow ACF for 1985-1990 Conditioned on Flows in 1983 and 1984

Upper Limit	1985	1986	1987	1988	1989	1990
500000	0	0	0	0	0	0
1000000	0	0	0	1	2	2
1500000	0	3	6	7	8	10
2000000	3	6	4	5	6	8
2500000	5	8	6	7	8	7
3000000	4	4	6	7	5	6
3500000	4	4	9	5	6	6
4000000	3	2	6	5	5	3
4500000	5	11	4	8	5	4
5000000	5	5	7	3	5	2
5500000	10	3	1	2	0	1
6000000	5	3	1	0	0	1
6500000	2	1	0	0	0	0
7000000	2	0	0	0	0	0
7500000	1	0	0	0	0	0
8000000	1	0	0	0	0	0
mean (ACFx10^6	4.37	3.47	3.15	2.95	2.72	2.56
std. dev. (ACFx10^6)	1.55	1.36	1.18	1.24	1.22	1.22

Table 6 Statistics and Observations of Explanatory Variables Used to Update the 1985 Annual Flow Based on January 1st Information.

Variable	Mean (1943-1984)	Observations 1985	Observations 1984
Total inflow from three rivers in Oct., Nov., and Dec.	337075.2	768900	1366580
Total lake precipitation in Oct., Nov., and Dec.	2.39	5.62	11.2
Snowpack on January 1st at Tony Grove, R.S.	4.2	8.7	15.3
Snowpack on January 1st at Parleys Canyon	7.1	12.6	18.0

Table 7 Histograms of 1985 Updated Flows (ACF) Using Information up to January 1st, 1985

Upper Limit	Number of Occurrences
5000000.	0
10000000.	0
15000000.	0
20000000.	1
25000000.	5
30000000.	4
35000000.	7
40000000.	10
45000000.	11
50000000.	5
55000000.	4
60000000.	2
65000000.	0
70000000.	1
75000000.	0
80000000.	0
Mean (ACFx10^6)	3.87
Std. Dev. (ACFx 10^6)	1.03

Table 8 Distributions of Predicted Future Annual Peak Weighted Average Elevations — 50 Simulations

	Starting Oct 1, 1984						Starting Jan 1, 1985
Lower Limit	1985	1986	1987	1988	1989	1990	1985
4200	0	0	0	0	0	0	0
4201	0	0	0	0	0	0	0
4202	0	0	0	0	0	0	0
4203	0	0	0	0	0	2	0
4204	0	0	0	0	1	1	0
4205	0	0	0	4	2	2	0
4206	0	0	4	1	4	4	0
4207	0	2	3	1	4	2	0
4208	7	8	3	3	0	1	0
4209	10	4	3	4	2	1	11
4210	5	3	5	3	5	5	21
4211	15	5	4	5	3	4	12
4212	7	5	2	1	3	5	5
4213	4	9	3	3	4	2	1
4214	2	7	9	9	2	1	0
4215	0	3	5	3	6	6	0
4216	0	4	2	4	3	2	0
4217	0	0	5	4	2	2	0
4218	0	0	2	2	4	4	0
4219	0	0	0	3	3	4	0
4220	0	0	0	0	2	0	0
4221	0	0	0	0	0	2	0
4222	0	0	0	0	0	0	0
50% Probability	4211.4	4212.8	4213.9	4213.9	4213.4	4212.5	4210.6
10% Probability	4213.1	4215.8	4217.4	4218.1	4219.0	4219.3	4212.3

ISSUES ASSOCIATED WITH STOCHASTIC MODELING OF GREAT SALT LAKE LEVELS FOR PLANNING PURPOSES

by David S. Bowles and L. Douglas James

INTRODUCTION

The rising level of the Great Salt Lake is having major impacts on the economy and on governments in Utah. Decision makers face uncertainties that scientific information can reduce. Principles from the physical sciences (meteorology, geology, hydrology, and physics), measured data (since 1843), and inferred information on prehistoric events must be combined with economic assessments. This paper presents the results of studies on probabilistic lake level forecasting and discusses issues of model reliability. A companion paper (James and Bowles 1985) describes how the probabilities can be used to optimize public lake-level-control and private lake-level-response investments. In combination, the papers identify some issues in model theory and data reliability to stimulate group discussion on research needs and management alternatives.

High precipitation, record river flows, and low evaporations have caused the lake water surface to rise more than twice as fast in the last two years as ever previously observed in 140 years of record. The south arm rose in successive annual highs to 4200.85 ft msl (June 1982), 4205.00 (July 1983), and 4209.25 (July 1984). Extrapolation suggests a peak in July 1985 of about 4211.0.

These rising levels are severely impacting private and public properties. Rough damage estimates are $350 million in 1985 and a $3.6 billion total by 2050. In the private sector, the mineral industry, railroads, and recreational enterprises are suffering major damages. Industrial, commercial, residential, and agricultural development are seriously threatened. In the public sector, federal, state, and local governments are experiencing large losses to roads and highways, waterfowl and related wildlife areas, and recreation facilities. Major expenditures are underway to protect several wastewater treatment plants and the Salt Lake International Airport.

METHODOLOGICAL FRAMEWORK

Water storage in the Great Salt Lake responds to precipitation, surface and groundwater inflows and evaporation outflows and changes water levels in a manner determined by the geometry of the lake bed. We do not know future inflows and outflows; we know past averages, variabilities, and serial and spatial correlations. We cannot forecast a specific stage for a future year; we can specify risks with probability distributions.

The methodological framework used four computer models. First, techniques from stochastic hydrology were used to generate 1000 series of annual sets of inflows and outflows. Second, these sequences were introduced into a water balance model to simulate 1000 sequences of lake levels. The distribution of simulated lake levels describes the risk in a particular year. People may respond to the risk by doing nothing and being inundated, evacuating, or investing to protect themselves. In the third model, these choices are simulated to translate each sequence of levels into a sequence of damages. The damages can be reduced by a project to control lake levels (pumping to the west desert, water development for increased consumptive use, etc.) or protect property (levees). The fourth model estimates the reduction in damages for each year of the sequence achieved by a specified project. The 1000 present worths of the reductions give a mean and a probability distribution of project benefits.

A multivariate first order autoregressive model was used to generate annual inflows and evaporations and coupled with water balance and damage simulation models by James et al. (1979). Later, James et al. (1981) corrected and updated their data file and reevaluated, confirmed, and recalibrated the selected stochastic model. However, both studies used data collected since 1937 when all the principal series were measured. Consequently, they do not reflect the high flows before 1880 and substantially underestimate probable lake levels. These earlier data were later approximated and combined with the recent very high flows to simulate the lake level sequences reported by James et al. (1984) with the methodology detailed in Bowles et al. (1985). These four reports supply the information for the presentation to follow.

HYDROLOGIC DATA

General Information

The Great Salt Lake was split into north and south arms in 1959 by construction by the Southern Pacific Railroad of a semipervious railroad causeway. The south arm receives most of the runoff (principally from the Bear, Weber, and Jordan Rivers). It has about twice the area, a higher water surface, and a lower salt concentration. The rapid rise over the last two years is plotted on Figure 1. The shallow lake expands or contracts over a large surface area (Table 1). At about 4217, the lake spreads onto the Western Desert and increases its surface area by about 650,000 acres.

Available Records

Reliable measurements began about 1875 for precipitation, 1876 for lake levels, 1890 for river flow, and 1937 for evaporation. Approximate lake level estimates and scattered precipitation data go to 1847. Temperature data go back to 1875 and were used (Hargreaves 1977) to estimate evaporation totals between 1875 and 1937 and during winter months. James et al. (1984) describe reconstruction of the four series from scattered lake level and precipitation records for the earlier years and correlations based on later data. The reconstructed tabulation shows eight of the ten largest river inflows to be in the very wet pairs shown in Table 2. The other two years were 1909 (4,060,000 acre-feet) and 1852 (3,470,000 acre-feet).

The historical sequences that could be reconstructed from reliable measured data were:

1. The combined total flows since 1890 for the Bear, Weber and Jordan Rivers (with releases from Willard Bay added since its construction in 1965), modified to reflect current land and water use practices (James et al. 1979) and referred to as "present modified flows."

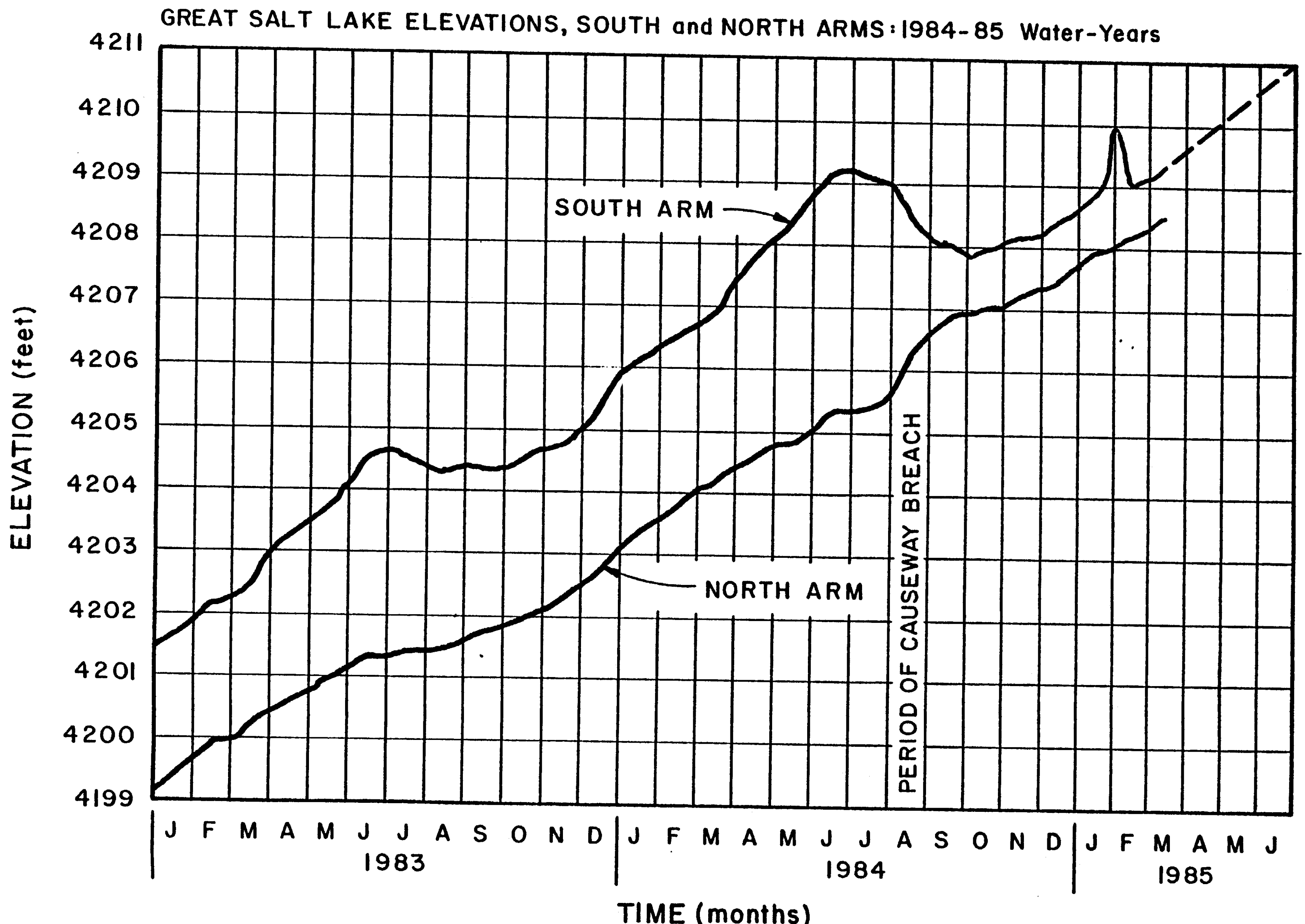

Figure 1. Great Salt Lake water levels during recent period of rapid rise, causeway breached in early August 1984.

Table 1. Physical data for the Great Salt Lake.

Water Surface Elevation (feet)[b]	Surface Area (1000 acres)	Storage Volume (1000 acre-feet)	Average Lake Evaporation (1000 AF/year)[c]	Mean Salt Concentration (percent)[a]
4190	564	7,868	1419	27.5
4191.35-L	587	8,570	1470	27.5
4195	720	11,002	1810	27.5
4200	1079	15,370	2908	22.5
4205	1251	21,277	3648	16.3
4210	1490	27,607	4550	12.5
4211.60-H	1556	30,057	4800	11.4
4216	2227	38,671	6900	8.9

[a]The saturation level of 27.5 percent is reached when the lake volume drops below 12,575,000 acre-feet at an average surface elevation of 4197. Actually, saturation occurs sooner in the brines of the North Arm.

[b]L = historic low - 1963, H = historic high - 1873

[c]Evaporation during an average year is an index of the inflow required to maintain the given water surface elevation.

Table 2. Data for four wettest 2-year periods.

Year	River Flow (acre-feet)	Rise (feet)	High Level (msl)
1983-84	11,946,000	8.4	4209.3
1871-72	8,430,000	0.8	4211.4
1921-22	7,480,000	2.4	4204.4
1862-63	6,670,000	3.0	4204.0

2. Precipitation on the lake surface based on Theissen polygons for weighting records kept since 1875 at Salt Lake City, Ogden, and Corinne. Amounts are adjusted downwards to reflect the lower average found over the lake for 1920 through 1929 from a closer network of gages at Kelton, Corinne, Farmington, Lakepoint, and Midlake.

3. Fresh water evaporation based on the pan data collected at the Bear River Bird Refuge beginning in 1937. Amounts back to 1875 were estimated from temperature data.

<u>Data Reliability Periods</u>

The 133 years of measured and reconstructed data fall into the following periods according to major changes in data availability:

1. 1938-1983. Period of Measured Data. The only data used that were not measured during this period were flows from 1938 through 1943 for the Jordan River and amount to less than 10 percent of the total surface inflow including precipitation.

2. 1890-1937. Period with Data Reconstructed from Independent Measurements. The Bear River flow and lake precipitation are measured, but some Weber (1890-1907) and all Jordan flows are estimated by correlation with the Bear. Evaporation is estimated from temperatures.

3. 1851-1889. Period with Data Reconstructed from Lake Stages. Precipitation and temperature data were used for 1875-1889, and scattered precipitation data were found for earlier years, but most of the data were reconstructed from correlations with lake volume changes based on period of measured data.

4. 1890-1983. Period of Independent Data.

5. 1851-1937. Period of Reconstructed Data.

6. 1851-1983. Total Historical Period.

Statistics for the evaporation, precipitation, and streamflow time series are compared on Table 3. Both precipitation and streamflow are highest before 1889, and the streamflow decreases period by period. The trend suggests that the modification used to convert historical flows to present land use conditions be reexamined. The 1851-1983 streamflow series has a significantly higher mean than does the series since 1938 at the 5 percent level.

<u>Anomalies in the Record</u>

The historical series contain two major anomalies, extraordinarily high river flows from 1862 through 1873 and in 1982-84. The earlier period was long enough to test for the statistical significance of the differences in the mean and standard deviation of its flows with those in two longer periods (1873-1983 and 1938-1983) (see Table 4). Both tests show the mean flows for 1862 to 1873 to be significantly higher at the 5 percent level; however, the standard deviations were not significantly different. The nonhomogeneity affects the mean but not the variance. Also, the precipitation on the lake did not test significantly higher. This may have been due to the poor quality of these early precipitation records or could result from some physical cause such as additional streamflow caused by a lake effect on precipitation in the mountains to the east or additional ground and surface inflow from the west during wet years.

The catchment to the west discharged about 300,000 acre-feet of surface runoff into the Great Salt Lake in 1983 after being dry since 1909. Water again flowed in 1984, but the higher lake level submerged the railroad culvert used

Table 3. Statistics for lake inflow sequences by data reliability period.

Period	Lake Evaporation (E) (ins) Mean	Standard Deviation[1]	Hurst Coefficient	Lake Precipitation (P) (ins) Mean	Standard Deviation[1]	Hurst Coefficient	Combined Streamflows (Q) (ac-ft) Mean	Standard Deviation[1]	Hurst Coefficient
1 1938-1983[2]	60.91	4.49	.66	10.76	2.37	.64	1,774,458	819,192	.63
2 1890-1937	61.55	4.22	.59	10.13	2.15	.51	1,988,506	837,218	.72
3 1851-1889	60.86	4.24	.69	10.80	3.00	.72	2,167,359	980,961	.76
4 1890-1983	61.24	4.34	.44	10.44	2.28	.48	1,883,760	826,542	.78
5 1851-1937	61.25	4.23	.50	10.43	2.58	.58	2,068,682	900,664	.68
6 1851-1983	61.13	4.31	.50	10.54	2.51	.49	1,966,921	877,976	.73

[1]Fiering corrected standard deviation (Fiering 1963).

[2]The lag 1 autocorrelation coefficients for this period were 0.385 for evaporation, 0.108 for precipitation, and 0.576 for streamflow. They were not estimated for the reconstructed data because the correlations were used in the reconstruction.

Table 4. Tests for the significance in the difference between the mean streamflow during a high flow period (1862-1872) and the means during two longer periods.

Period	1	2	3
Dates	1862 to 1872	1873 to 1983	1938 to 1983
$\bar{X}$	3,137,000	1,898,968	1,803,807
S	794,742	798,679	819,192
n	11	111	46
$(\bar{X}_1-\bar{X}_2$ or 3)		1,238,032	1,333,193
$w = S^2/n$	5.742×10^{10}	5.747×10^{9}	1.981×10^{10}
$t_{.05}$	2.20	1.98	2.01
$w_1t_1 + w_2t_2$ (or 3)		1.377×10^{11}	1.661×10^{11}
$(w_1t_1 + w_2t_2)/(w_1 + w_2)$ (or 3)		2.180	2.151
T_{crit}		4.93	4.80
Inference	Significantly different means	Significantly different means	Significantly different means

to estimate the flow, and no totals are available. The flow in 1983 came from north of the Southern Pacific Railway; wetter conditions may introduce flow from the much larger area further south.

PROBABILISTIC HYDROLOGIC MODELING

Alternative Approaches

The long record of annual lake level highs could not be used directly to estimate lake level probabilities because the high serial correlation (lag one = 0.98, Hurst = 1.08) violates the assumption of independent events. A second possible approach, direct stochastic generation of lake level sequences, would tax our ability to calibrate a representative stochastic model because of the large number of lags required to match the high lake level persistence (James et al. 1979). Furthermore, a lake level sequence provides no way to examine the effectiveness of measures changing inflows or outflows as control methods. These and other approaches are evaluated in Table 5.

Table 5. Alternative probabilistic modeling approaches.

Model	Assessment
1. Direct frequency analysis of historic lake levels	Ignores interdependence of events
2 Generation of lake level sequences	Lag and Hurst coefficients are outside the range that can be preserved; cannot be directly applied in evaluating control alternatives
3. Generation of algebraic sums of annual flows	Gives composite events that add doubt when extrapolating
4. Generation of multivariate, multilag series	No available model
5. Generation of multivariate, single lag series	Used in this study and generally found to give good results but may not adequately preserve long term persistence
6. Generation of principal variate, multiple lag series. Use correlation to estimate other variables	Used by Bras et al. (1985). Poor job of preserving serial correlations in secondary variates
7. Seasonal or monthly generation	Extra work seems unwarranted where annual peak is desired; no data base for incorporating large 19th century flows

HYDROLOGIC TIME SERIES MODELING

A variety of stochastic models are available for generating hydrologic time series. The selection can be approached in a logical and systematic manner (Box and Jenkins 1970) by an iterative approach (James et al. 1981), with five steps (Figure 2). 1) identification of the water resources system and model composition (e.g., univariate or multivariate, annual or seasonal); 2) choice of model type--short term or long term persistence model (e.g., autoregressive or fractional Gaussian noise); 3) identification of model form (e.g., order of the autoregressive model); 4) parameter estimation; and 5) model performance evaluation.

Statistics to Be Preserved

Model building starts by analyzing the available records to understand the statistical structure which the stochastic model should preserve. The structure includes:

1) Distributional properties of each flow series (mean, variance, skew, and probability distribution)

2) Serial correlation in each flow series (autocorrelogram)

3) Cross-correlations among the flow series (cross-correlation matrices including lagged cross correlations)

Three possible assumptions in applying the computed structure are:

1. The historical record provides homogeneous data, and the computed statistics represent long term conditions.

2. The historical record is too short, and the computed statistics should be modified to represent long-term conditions, perhaps based on tree ring or geomorphic data.

3. The historical record contains nonhomogeneities caused by exogenous events (volcanic eruptions, sunspot cycles, and planetary tides are mentioned), and the statistics should be adjusted and perhaps varied over the simulation period to represent patterns in these events.

The results obtained with the study roughly matched the indications of the tree ring (James et al. 1979) and geomorphic data (Utah Council on Science and Technology 1984). Possible exogenous influences were explored, but no statistically significant relationships were found. Thus no adjustments were made.

Application to the Great Salt Lake

James et al. (1979) tried several stochastic models and selected a multivariate AR(1) (equivalent to a multivariate ARMA (1,0)) model (Box and Jenkins 1970) fitted to the period 1937-1977. By using only this period, the calibration avoided the loss of the random component in the cross-correlation used to reconstruct series. However, by ignoring the earlier wet period, the probabilities of high lake levels were underestimated.

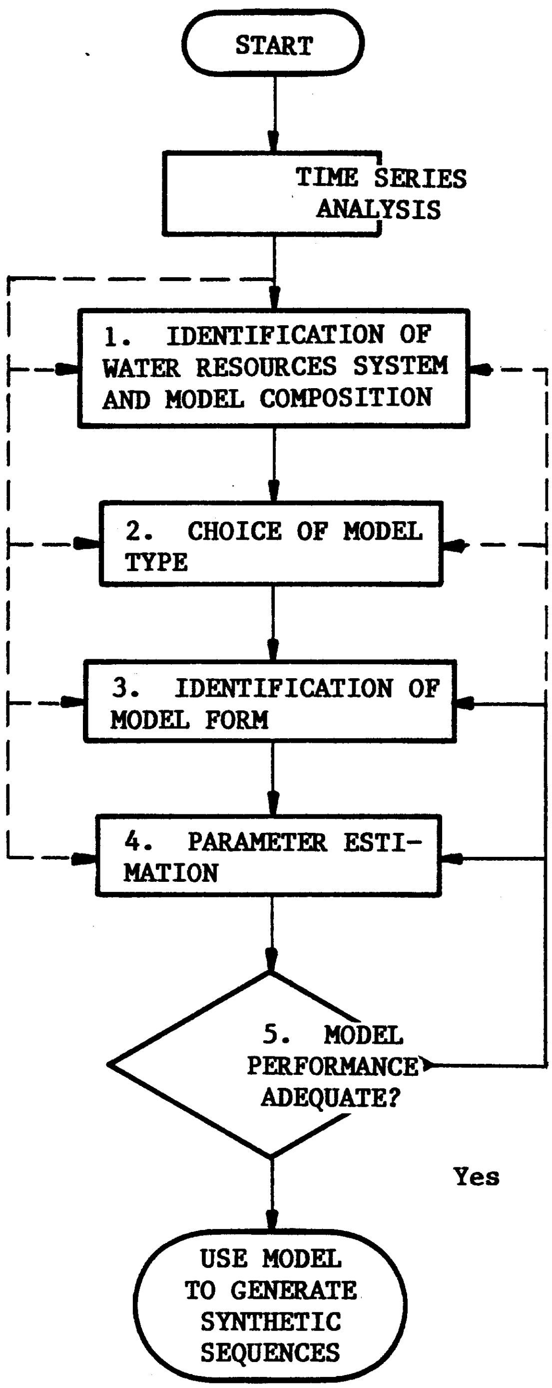

Figure 2. Systematic procedure for stochastic modeling of hydrologic time series (from James, Bowles, and Kottegoda 1981).

The following study, therefore, sought to consider these higher flows and yet preserve the cross-correlation structure. The goal in calibrating the multivariate model was to preserve the mean and standard deviation of the 1851-1983 series and the first two correlation matrices (M_0 and M_1) of the 1938-1983 series. Table 6 tabulates the lag zero (M_0), lag one (M_1), and lag two (M_2) correlation matrices for the shorter series. A two year flow-lag effect is the only significant cross correlation in the M2 matrix. The mean and standard deviation from the longer series was chosen to do a better job of representing hydrologic conditions observed over the last 133 years. The data series were normalized by applying the relationship:

$$Z_t = \frac{\ln (X_t - \beta) - \mu}{\sigma} \qquad (1)$$

A two-parameter log normal transform (i.e., $\beta = 0$) was used for precipitation and streamflow, and a three-parameter log normal transform (with 3.67 ft as the value of β) was used for evaporation.

Model Calibration

The significant relationships within and among the three series were preserved by calibrating a first order autogressive model of the form:

$$A_t = A\ Z_{t-1} + B\ E_t \qquad (2)$$

where the values for the 3-by-3 matrices, A and B, are given in Table 7, the A_t are the values of the three time series in consecutive years, and the E_t are normally-distributed random errors of zero mean and unit standard deivation (N(0,1)). Parameter estimation followed methods described by Salas et al. (1980).

Model Performance Evaluation

Model performance is judged by the preservation of historical statistics and the independence of residuals calculated by subtracting the results with the stochastic model from the historic record. Inadequate performance may require changing the values of the model parameters or, if this does not work, the model form. If inadequate model performance persists, it may be necessary to select an alternative model or simplify the model composition.

For comparing generated with historical statistics, 100 sets of the three flow series were generated. Each set corresponded to the length of the 1938-1983 historical period and began from 1938 observed values. For comparison, the means were also based on the 1938-1983 period. The comparisons (Table 6) show all the statistics for the generated flows to be within one standard deviation of the mean of the historical statistics, thus indicating a good preservation.

Another test was to examine the high flow periods. Fifty series were generated, using 1983 initial conditions, with means estimated from the 1938-1983 period. Six generated values exceeded the high 1983 inflow level of 5.1 million acre-feet. Eleven generated 25-year series exceeded the 1872 high. It was concluded that the model was capable of generating inflow sequences larger than those during 1862-1873.

Table 6. Comparison of statistics for historical and generated sequences.

Historical statistics based on the period 1938-1983
Average for 100 series generated for same length from 1938 initial conditions
Standard deviation of 100 generated series

		Lake Evaporation (E)	Lake Precipitation (P)	Combined Streamflow (Q)
a) Mean				
Hist. Stats		60.91	10.76	1774458*
Aver. of Gen. Means		61.18	10.79	1775080
St. Dev. of Gen. Means		1.11	0.44	204931
b) Standard Deviation				
Hist. Stats		4.49	2.37	819192*
Aver. of Gen. Std. Dev.		4.39	2.26	757244
St. Dev. of Gen. Std. Dev.		0.56	0.32	190209
c) Hurst coefficient				
Hist. Corr.		0.66	0.64	0.63
Aver. of Gen. Hurst coef.		0.68	0.49	0.65
St. Dev. of Gen. Hurst coef.		0.09	0.15	0.13
d) Lag-zero cross-correlation matrix				
Hist. Corr.	E	1.000	-0.622	-0.449
Aver. of Gen. Lag-zero cross-corr.		1.000	-0.605	-0.433
St. Dev. of Gen. Lag-zero cross-corr.		0.000	0.137	0.172
Hist. Corr.	P		1.000	0.614
Aver. of Gen. Lag-zero cross-corr.			1.000	0.593
St. Dev. of Gen. Lag-zero cross-corr.			0.000	0.104
Hist. Corr.	Q			1.000
Aver. of Gen. Lag-zero cross-corr.				1.000
St. Dev. of Gen. Lag-zero cross-corr.				0.000
e) Lag-one cross-correlation matrix				
Hist. Corr.	E	0.378	-0.037	-0.149
Aver. of Gen. Lag-one cross-corr.		0.326	0.008	-0.101
St. Dev. of Gen. Lag-one cross-corr.		0.166	0.185	0.208
Hist. Corr.	P	0.005	0.090	0.167
Aver. of Gen. Lag-one cross-corr.		0.035	0.048	0.110
St. Dev. of Gen. Lag-one cross-corr.		0.158	0.182	0.176
Hist. Corr.	Q	-0.459	0.614	0.589
Aver. of Gen. Lag-one cross-corr.		-0.446	0.589	0.519
St. Dev. of Gen. Lag-one cross-corr.		0.169	0.118	0.138
f) Lag-two cross-correlation matrix				
Hist. Corr.	E	0.258	-0.017	-0.040
Aver. of Gen. Lag-two cross-corr.		0.211	-0.040	-0.073
St. Dev. of Gen. Lag-two cross-corr.		0.160	0.176	0.208
Hist. Corr.	P	-0.124	0.198	0.186
Aver. of Gen. Lag-two cross-corr.		-0.013	0.075	0.071
St. Dev. of Gen. Lag-two cross-corr.		0.182	0.162	0.166
Hist. Corr.	Q	-0.232	0.181	0.362
Aver. of Gen. Lag-two cross-corr.		-0.144	0.177	0.188
St. Dev. of Gen. Lag-two cross-corr.		0.195	0.184	0.203

Underlined historical data are significant at the 5 percent level.

*This table compares how well the model preserves the various statistics. The actual generation preserved the 1851-1983 mean of 1,966,921 and standard deviation of 877,976 (Table 3).

Table 7. Matrix values in the calibrated model (Equation 2).

A Matrix			B Matrix		
0.56284	0.37749	-0.12772	0.88619	0	0
0.12734	0.07246	0.18434	-0.72954	0.65512	0
-0.07088	0.37300	0.30935	-0.41909	0.34741	0.51517

If the selected stochastic model fits the historical record well, the residuals are independent and normally distributed with zero mean and unit variance. Also, there should be no significant cross-correlations between the three residual series corresponding to the three inflow variables. Residuals series were calculated, and no significant autocorrelation structure was inferred at the 5 percent level for lake evaporation or lake precipitation. However, the streamflow residuals showed significant autocorrelation during periods which included the 1862-1873 anomaly. This suggests that the generated streamflows were showing less autocorrelation during the late 1800s than was observed. When tested at the 1 percent level, the autocorrelation was not significant provided the mean and standard deviation from the entire record, 1851-1983, were used. These observations support use of the 1851-1983 period for estimating the means.

Estimates of the means and standard deviations for the residuals series are presented in Table 8. Again the residuals for evaporation and precipitation indicate a satisfactory model. However, statistically significant differences exist for the combined streamflow series. Nevertheless, overall, the performance was judged acceptable.

Lake Water Balance Model

The sequences for the three generated flows were introduced into an annual lake water balance model developed by the Utah Division of Water Resources (1974). Two smaller sequences (ungaged streamflow and groundwater flow called S_t and G_t) could not be generated because no data have been measured. Runs with the lake water balance model to match historical stages gave as relationships for estimating these two variables:

$$S_t = 0.2955*P_t*A_t - 0.0226*E_t*A_t \tag{3}$$

$$G_t = 0.015*Q_t + 0.015*Q_{t-1} + 0.015*Q_{t-2} \tag{4}$$

The forms Equations 3 and 4 assume that groundwater inflow is time lagged and that ungaged streamflow from small catchments varies with precipitation and evaporation (antecedent moisture). Use of these relationships reduces the variability in the stochastic model only slightly because the numbers are small.

The generated evaporation sequence was converted from pan to freshwater lake evaporation by multiplication by 0.797 and then adjusted for salinity effects by using:

Table 8. Means and standard deviations of residuals of multivariate AR(1) model[1] for 1851-1983 period. (Underlined values are statistically different from the theoretical values at the 5 percent level of significance.)

	Lake Evaporation	Lake Precipitation	Combined Streamflow
a) Means and variances estimated from 1851-1983 period[2]			
Mean	-0.014	-0.002	0.022
Standard Deviation	1.12	0.87	1.64
b) Mean and variance estimated from 1938-1983 period[2]			
Mean	0.094	-0.117	0.57
Standard Deviation	1.11	0.96	1.79

[1]Theoretical values of mean and standard deviation are zero and unity, respectively.

[2]A and B parameter matrices estimated from 1938-1983 period.

$$E_{sl} = E_f\,(1.0 - 0.00833*C) \tag{5}$$

where C = the mean lake salinity as a percentage. Both evaporation and precipitation were generated as annual totals per unit area and multiplied by the area of the lake.

Estimation of Annual Peak

The water balance modeling was done by water year. Estimation of the annual peaks required separations of the precipitation, evaporation, and flow that occur before the peak in a water year from those afterwards. Data were assembled on the dates of the peak lake elevations (Table 9), and the three fractions were correlated with the annual totals for the flow, precipitation, and evaporation.

Simulation of Sequences

Beginning from the antecedent conditions at the end of the 1983 water year, 1000 event sequences, each covering the period to 2050, were generated and entered into the lake water balance model to generate 1000 lake level sequences. Each sequence gives an equally probable scenario of future conditions; and sorting to determine the nth highest level gives an estimate of the

Table 9. Relationship between date of recorded lake high and river inflow (1957-1983).

Date of High	Number of Years	Mean River Inflow (million AF)
March 15	1	1.07
April 1	2	0.72
April 15	1	1.58
May 1	2	1.34
May 15	5	2.11
June 1	5	2.15
June 15	6	1.78
July 1	4	2.70

n/10 percent event. Table 10 shows probability distributions for various years in the future, and Table 11 shows the probabilities of the lake reaching various elevations at least once by the years indicated. As an example of how to read Table 10, there was, as of October 1, 1984, one chance in 100 (to be expected should the winter bring the 100-year precipitation) of the lake reaching 4213.1 in 1985 and a 53 percent chance of the lake reaching 4212 at least once by 2010.

APPROXIMATE FORECASTING METHOD

One begins a water year with a great deal of uncertainty as to high lake level that will be reached the following summer. However, as the year progresses, one can improve the estimate from updated information. Precipitation measurements provide an excellent and easily obtained index, and the Great Salt Lake precipitation data show relatively low month-to-month correlations. The historical frequency distribution of annual precipitation at the Salt Lake Airport (1940-1983) was plotted and used to update estimates of wetness frequency as shown on Table 12.

The cumulative departures show 1984 to have been wetter than normal from the start, and particularly heavy rains in December and April caused 5 percent annual precipitation event. The 1984 distribution (similar to Table 10 but for the previous year) indicated a level of 4209.0 for this event. The recorded annual high on July 1 was 4209.25, a relatively minor difference. As of March, the method predicts a 1985 peak lake level of 4211.0.

STRUCTURAL ALTERNATIVES

The Utah Division of State Lands and Forestry is responsible for general management of the lake. The Division formulated a contingency management plan (Utah Division of State Lands and Forestry 1983), and the Utah Division of Water Resources (1984) is exploring structural lake level control methods. Two alternatives that would affect the levels over the lake as a whole are:

1. Pumping lake water into western desert storage (Eckhoff, Watson, and Preator Engineering 1983). The ponds would have a storage capacity of 1.33 maf over 450,000 acres and add 1.06 maf to the annual lake evaporation. The

Table 10. Probability distributions of annual high levels of the Great Salt Lake given 1851-1984 data, adjusted to reflect 1965 land use, assuming no predictable cyclic weather patterns, and using a water balance analysis.

Year	Probabilities of Exceeding			Mean	Probabilities of Dropping Below		
	0.01	0.10	0.25		0.25	0.10	0.01
1984	4209.2	4209.2	4209.2	4209.2	4209.2	4209.2	4209.2
1985	4213.1	4212.0	4211.3	4210.7	4210.2	4209.8	4209.2
1986	4216.2	4213.8	4212.6	4211.3	4210.2	4209.1	4207.1
1987	4217.7	4214.6	4213.0	4211.3	4209.5	4208.1	4206.1
1988	4218.5	4214.7	4213.0	4210.8	4208.4	4206.8	4204.5
1989	4219.0	4214.5	4212.6	4209.9	4207.3	4205.3	4202.7
1990	4218.4	4214.3	4212.0	4208.8	4206.1	4204.0	4201.0
1995	4216.2	4211.9	4208.2	4204.4	4201.2	4198.6	4195.0
2000	4214.6	4209.7	4205.4	4201.8	4198.6	4196.1	4191.1
2010	4213.9	4206.9	4202.8	4199.7	4196.7	4194.1	4188.2
2020	4212.9	4206.1	4202.5	4199.3	4196.2	4193.6	4187.4
2030	4212.5	4206.3	4202.5	4199.2	4196.0	4193.0	4186.5
2040	4213.2	4206.0	4202.4	4199.2	4196.3	4193.6	4186.5
2050	4213.8	4206.0	4202.4	4199.2	4196.3	4193.4	4186.3

Table 11. Probability of the annual high level of the Great Salt Lake exceeding given stage at least once by the given year based on the above tabulation and assuming independent annual events.

Year	4210	4211	4212	4214	4216	4218	4220
1985	0.839	0.386	0.094	0.001	0.000	0.000	0.000
1986	0.864	0.582	0.378	0.083	0.014	0.001	0.000
1987	0.872	0.623	0.441	0.141	0.036	0.007	0.002
1988	0.876	0.643	0.467	0.182	0.048	0.015	0.002
1989	0.880	0.649	0.477	0.190	0.058	0.016	0.005
1990	0.880	0.654	0.483	0.206	0.066	0.017	0.006
1995	0.889	0.675	0.506	0.224	0.082	0.024	0.006
2000	0.893	0.680	0.512	0.233	0.086	0.027	0.007
2010	0.900	0.697	0.530	0.244	0.093	0.028	0.009
2020	0.907	0.710	0.538	0.253	0.098	0.030	0.010
2030	0.913	0.721	0.551	0.260	0.101	0.031	0.011
2040	0.915	0.727	0.563	0.271	0.108	0.033	0.012
2050	0.916	0.734	0.569	0.278	0.113	0.035	0.013

Table 12. Approximate tabulation for monthly updating for estimating a probability for water year 1984.

Month	Mean Airport Precipitation	1983-84 Airport Precipitation	1983-84 Departure	Cumulative Departure	Precipitation Frequency	Most Probable 1984 High
						4206.9
October	1.33	1.62	+0.29			
				+0.29	45%	4207.1
November	1.27	2.23	+0.96			
				+1.25	33%	4207.5
December	1.36	4.37	+3.01			
				+4.26	10%	4208.4
January	1.32	0.50	-0.82			
				+3.44	15%	4208.2
February	1.23	0.95	-0.28			
				+3.16	18%	4208.1
March	1.81	0.95	-0.86			
				+2.30	23%	4207.9
April	2.08	4.43	+2.35			
				+4.65	9%	4208.5
May	1.59	1.98	+0.39			
				+5.04	8%	4208.6
June	1.06	1.86	+0.80			
				+5.84	5%	4209.0
Sum	13.05	18.89				

operating assumption is that pumping will begin when the lake level passes 4202.0 feet msl and that the storage in the desert will reduce lake levels during the period of rapid rise. Afterwards, the pond surface adds evaporation. A flow of 500 cfs is maintained back to the lake to preserve the salt balance. As the lake recedes, all the water drains back to the lake.

2. Constructing upstream reservoirs and diverting up to the 390,000 acre-feet per year permitted by the Bear River Compact for consumptive use (Utah Division of Water Resources 1984).

The two control methods can be employed in various combinations and will take different amounts of time before they can be implemented. The fastest reasonable schedule is to begin pumping for 1987 and have additional reservoir storage in place by 1994. The effect of each scheme was simulated by incorporating their operation within the lake water balance model. Illustrative probabilities of the effects are shown in Table 13.

MODELING ISSUES

Several major issues should be considered in assessing the reliability of the hydrologic modeling presented above:

Table 13. Probability distributions of the expected effects on annual high levels of the Great Salt Lake with various management alternatives given the information available October 1, 1983.

Year	Probabilities of High Level Being Exceeded						
	0.990	0.900	0.750	0.500	0.250	0.100	0.010
P87 - West Desert Pumping in 1987							
1987	0	0.7	0.9	0.9	0.9	0.8	0.5
1990	0.3	0.6	1.1	1.7	3.2	3.6	2.9
1995	1.2	0.5	0.7	1.2	1.9	4.1	5.0
2000	0.5	0.6	0.5	0.8	1.3	3.0	5.4
2020	0.5	0.2	0.2	0.4	0.8	2.1	4.1
P87 + R94 - Pumping in 1987, 300,000 AF/yr Consumptive use in 1994							
1995	0.9	0.8	0.7	0.6	0.5	0.1	0.5
2000	3.0	1.9	1.8	1.3	1.0	0.4	1.2
2020	4.5	3.3	2.6	1.9	1.3	0.6	1.2

1. Representativeness of the Available Measurements. The data series should represent the sum of the surface and groundwater inflow to the lake, the precipitation on the lake, and the evaporation from the lake. Major potential problems in using historical measurements include a) the possibility of major flows entering from the west desert during wet years, b) variability between storm patterns over the lake and the precipitation indicated by gages to the east, c) the effect of salt content and mixing in the surface waters of the lake on evaporation rates and the variability of that mixing with weather patterns, d) errors in adjusting the gaged historical streamflow record for the effect of changes in consumptive use in the tributary basin with human development, e) poor information on groundwater inflow to the lake, and f) discontinuities in measured records caused by changes in gage locations or procedures.

2. Anomalies in the Historical Data. A statistical test described in the body of this paper indicated that the mean streamflow from 1862 through 1873 significantly differed from that for the balance of the record. A similar shift may have occurred in 1982. Ideally, one would expect these changes to have physical bases that can be identified and quantified for risk assessment. Such physical factors as atmospheric and oceanic circulation, volcanic eruptions, and solar weather should be carefully examined.

3. Preserving Historical Persistence. Terminal lakes invariably occur in drier climates exhibiting much greater variability and persistence in streamflow than is common of more humid climates. The result is that streamflows tend to be serially correlated over fairly long periods and much harder to model. Hydrologic forecasting is dealing with joint probabilities controlling extremely rare events.

4. Evaporation-Streamflow Feedback. The geographical setting of the Great Salt Lake, just upwind from a major mountain range that is the source

of most of its inflow, substantiates speculation that the high persistence found in the precipitation records in these mountains may be caused by the additional evaporation that occurs at higher lake levels.

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What General Circulation Models Say About the Effect of Equatorial Pacific Sea Surface Temperature on Midlatitude Winter Climate

J. E. Geisler
University of Utah, Department of Meteorology, Salt Lake City, UT 84112

1. Introduction

The broad question addressed in this presentation is that of the extent to which winter climate anomaly patterns over North America are linked to warm sea surface temperature anomalies (SSTAs) in the equatorial Pacific Ocean. The specific question addressed is that of the extent to which we can, through the use of a global atmospheric general circulation model (GCM), translate an observed warm equatorial Pacific SSTA into a North American winter climate pattern forecast. The term forecast is used here because observed warm equatorial Pacific SSTAs evolve rather slowly, on a time scale of months, thus providing essentially constant forcing for model predictions of monthly or even seasonally averaged climate anomaly patterns.

A warm SSTA appears in the eastern or central Pacific Ocean once every three to five years on the average. Popularly known as an El Niño event, the most recent of these episodes occurred in the winter of 1982-83. A variety of catastrophic meteorological phenomena, principally floods and droughts, occurring around the globe in tropical latitudes during 1982-83 were demonstrably linked to this warm SSTA. Midlatitude winter climate anomalies in 1982-83 were also attributed to this warm SSTA. We will argue in this paper that naturally occurring interannual variability of winter climate (so-called climatic noise) is so large as to obscure the cause-effect relationship between a midlatitude winter climate anomaly and the presence of a warm SSTA in the equatorial Pacific. We also discuss a recent finding that the generation by anomalous tropical precipitation of large-scale waves capable of propagating into the midlatitude winter hemisphere is more complicated than previously thought.

2. Observed Variability of Midlatitude Winter Climate

The term climate anomaly pattern is defined here as the difference between a particular atmospheric field averaged over a particular winter season and the same seasonally averaged field further averaged over an ensemble of winters. A field commonly examined in climate anomaly studies is the height of the 500 millibar (mb) pressure surface, situated on a average about 5.5 km above the ground in Northern Hemisphere winter. Fig. 1a shows the ensemble-average pattern of 120-day average (mid-November to mid-March) values of this field, the ensemble consisting of the 11 winters in the period from 1965-66 to 1975-76. These contours of 500 mb height are essentially streamlines of the flow on this pressure surface. The winds at this level blow from west to east around the globe in a meandering pattern of climatological-mean ridges and troughs.

Each of the 11 winters (which can be viewed as 120-day-average samples) that goes into this ensemble average is considerably different from the others. These differences are revealed in the standard deviation of the field, which is shown in Fig. 1b. This is seen to increase steadily with latitude, ranging from 70 m to 140 m over the United States. The maxima in the North Pacific and in the North Atlantic reflect the high degree of winter storminess there.

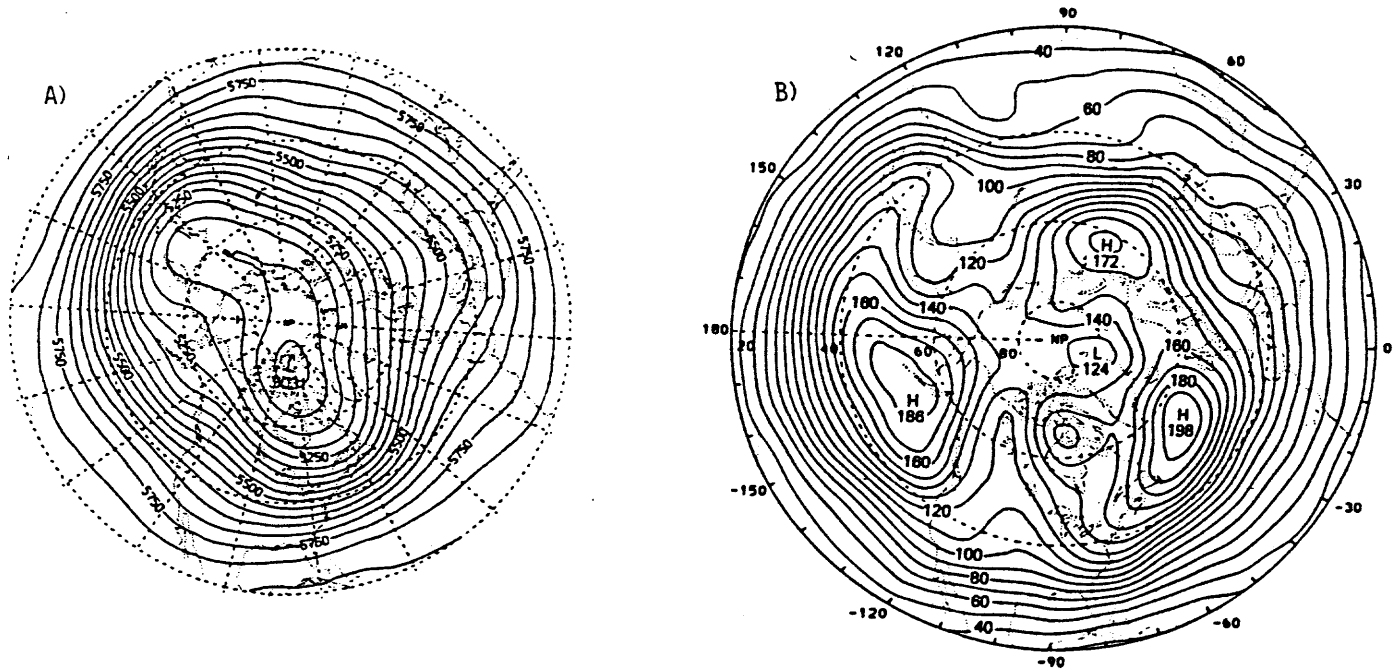

Fig. 1 Contours of a) observed long-term winter average height of the 500 mb surface (from Lau, et al., 1981) and b) standard deviation (interannual variability) of this height (from Lau et al., 1981). Units are meters.

3. Simulated Variability of Midlatitude Winter Climate

For winter climate experiments, we use the atmospheric general circulation model (GCM) of the National Center for Atmospheric Research. We operate it in the perpetual January mode, in which sea surface temperature, land surface albedo, and incoming solar radiation are held fixed at January-average values for a 1200-day run of the model. This record is then divided up into eight independent 90-day average samples of winter climate. The samples are rendered independent by the omission of 60-day lengths of the record between samples. The average of these eight samples of winter constitutes the ensemble average. Fig. 2a shows the ensemble-average 500 mb height-field map thus obtained. It compares favorably with the observed pattern that was shown in Fig. 1a.

Even though global SST is held constant, each of the eight winter samples in our simulation differs considerably from the others. As noted previously, the standard deviation of these eight sample mean fields is a measure of how much they differ. This statistic is shown in Fig. 2b. Comparison with the standard deviation of the observed 500 mb height field (Fig. 1b) shows that the inter-sample variability in the simulated fields is about the same size as in the observed fields, not only over North America, but also everywhere else in the Northern Hemisphere. This is a matter of some considerable significance. Inter-sample variability between the simulated winters arises entirely from internal atmospheric processes, while the observed inter-sample variability arises from internal atmospheric processes plus the interannual variation of global SST patterns and land surface albedo. The key fact illustrated by comparison of Fig. 1b and 2b is that the atmospheric system is so noisy, even when we consider a field time-averaged over the whole winter, that interannual changes in global SST

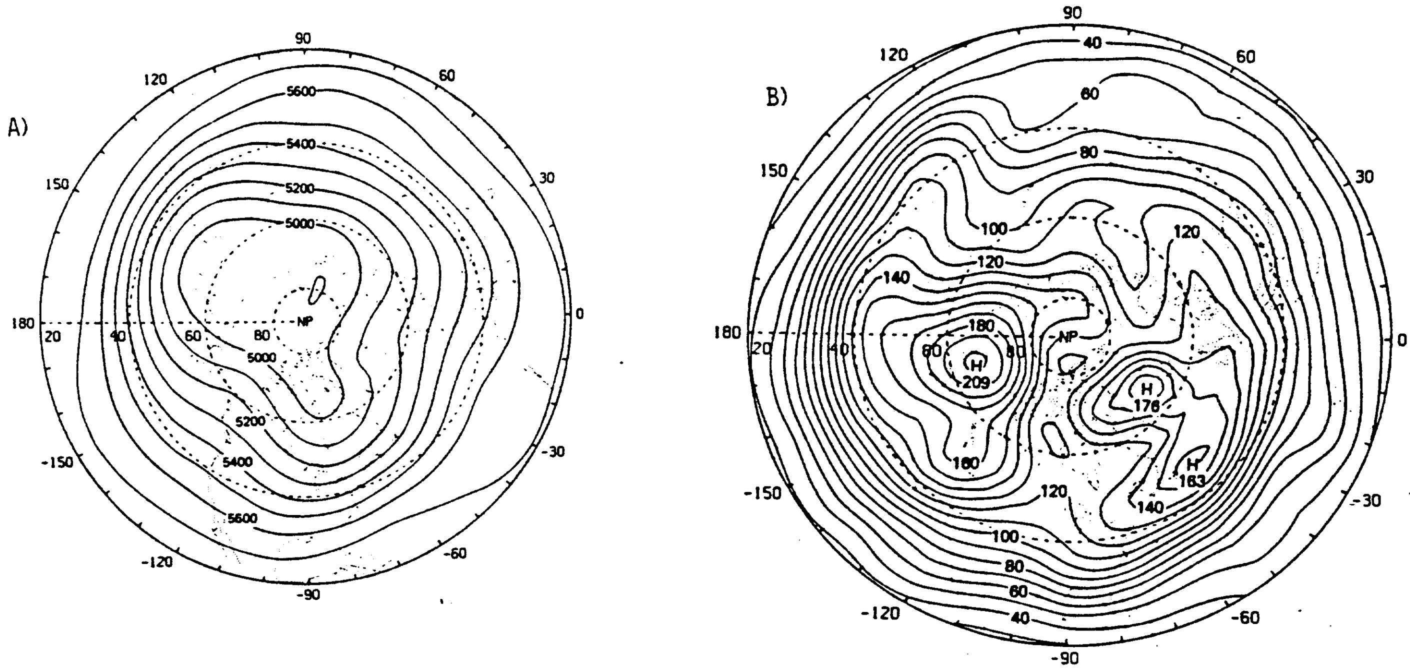

Fig. 2 Contours of a) GCM-simulated long-term winter average height of the 500 mb surface (from Pitcher et al., 1983) and b) the standard deviation (inter-sample variability) of this field (from Malone et al., 1984). Units are meters.

patterns seem to produce little or no extra variance. In particular, we note that three warm equatorial Pacific SSTA anomaly events that occurred during the 11-year observational record do not noticeably increase the observed standard deviation over the simulated (constant-ocean SST) standard deviation.

This comparison of standard deviations is of course a rather crude test, but it suggests that there is sufficient noise in the midlatitude winter hemisphere to obscure the biggest interannual global signal that we are aware of: the response of the atmosphere to the episodic appearance every three to five years of a warm SSTA in the equatorial Pacific.

The source of the natural variability (noise) in seasonal-average climate patterns in any atmospheric field (200 mb height, surface pressure, surface temperature, etc.) is the wide fluctuations in midlatitude winter weather accompanying the transient (time-scale of a few days) ridges and troughs spoken of in daily weather forecasts. It has been known for the last twenty years that the appearance, disappearance and movement of these systems cannot even in principle be predicted beyond a period of about 10 days. (In present practice, the predictability of these systems is limited to a few days and is far from perfect.) All of the standard deviation in Fig. 2b (that is, the differences between our eight 90-day average sample winters) is due to weather in the sense defined above. Since the 90-day averaging time is well beyond 10 days, the time-average fields in these samples are not predictable, hence the grounds for referring to the inter-sample variability exhibited by Fig. 2b as "noise." The comparison with Fig. 1b makes the point that much of the observed interannual variability of midlatitude winter climate is apparently also noise. Further analysis and assessment of these matters can be found in a paper by Madden (1976).

4. An SSTA Experiment

We showed in Fig. 2a the ensemble average of eight independent realizations of 90-day average winter climate, which we will refer to as climatology or as the control run. Fig. 3 shows an equatorial Pacific SSTA representative of the observed SSTA averaged over the central months of the winter of 1982-83. We have added this SSTA to the (global) SST pattern used for the control run. With this new SST, we have run the GCM for 1200 days in the perpetual January mode, thereby generating eight independent realizations of 90-day average winter climate in the presence of the 1982-83 winter equatorial Pacific SSTA. We have formed the ensemble average of these eight samples and subtracted climatology (that is, the control run) from it.

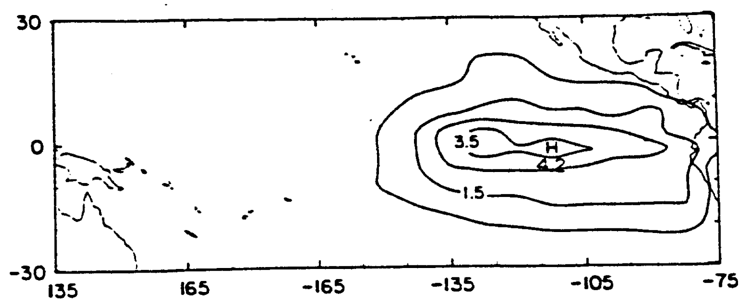

Fig. 3 Equatorial Pacific SSTA (in deg K) used in the anomaly experiment.

A detailed analysis of this and related GCM experiments is presented in Geisler et al. (1985). Here we show in Fig. 4a the resulting 200 mb height anomaly. There is present near 135°W in the equatorial Pacific a well-defined pair of anticyclones that are forced by a pronounced equatorial precipitation anomaly (not shown) that is situated between them, on the equator. This simulated feature is in good agreement with what was seen in satellite imagery of equatorial Pacific cloud patterns during this period. The largest midlatitude anomaly in Fig. 4a is the pattern over the Pacific and North American sector of the globe. As a measure of the statistical significance of this pattern we have computed the statistic

$$t = \left(\frac{N}{2\sigma}\right)^{1/2} \Delta h \quad .$$

where Δh is the 200 mb height anomaly as shown in Fig. 4a, σ is the standard deviation of 90-day average 200 mb heights (assumed equal in the control run and the SSTA run) and $N = 8$ is the number of samples. Fig. 4b shows contours of this statistic. We use the two-sided t-test to establish the significance level (probability of incorrectly rejecting the null hypothesis that $\Delta h = 0$). We find that, for the number of degrees of freedom $2(N-1) = 14$ in our case, $t = 2.15$ is the 5% significance level. The pattern over the midlatitude Pacific and over Northern Canada is significant at the 5% level. The pattern over the United States is not significant at the 5% level, except perhaps along the Gulf Coast.

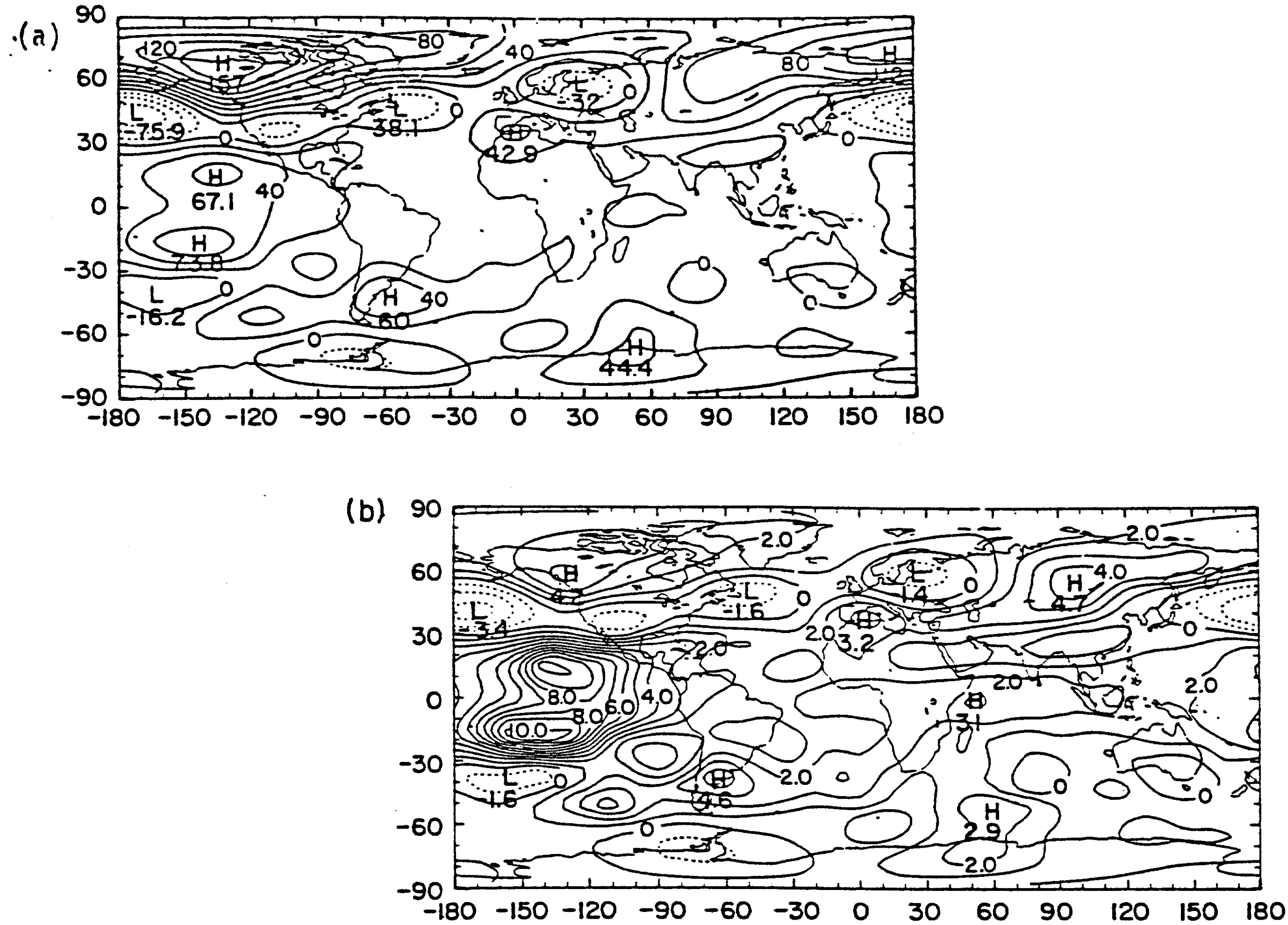

Fig. 4 a) The 200 mb height anomaly pattern obtained by subtracting the control run from the anomaly run and b) contours of the t-statistic for this field (from Geisler et al., 1985). Units in a) are meters.

To illustrate the inter-sample variability that underlies the marginal statistical significance of the ensemble-average pattern over the United States in Fig. 4a, we have assembled in Fig. 5 four of the eight samples that go into making up that ensemble-average. Shown in Fig. 6 is the observed 200 mb height anomaly for March 1983. It can be seen that the ensemble-average pattern (Fig. 4a) is not very representative of this observed pattern. What can also be discerned is that the spread of individual samples in the simulation (Fig. 5) is almost as big as the difference between the observed pattern and the ensemble-average simulated pattern.

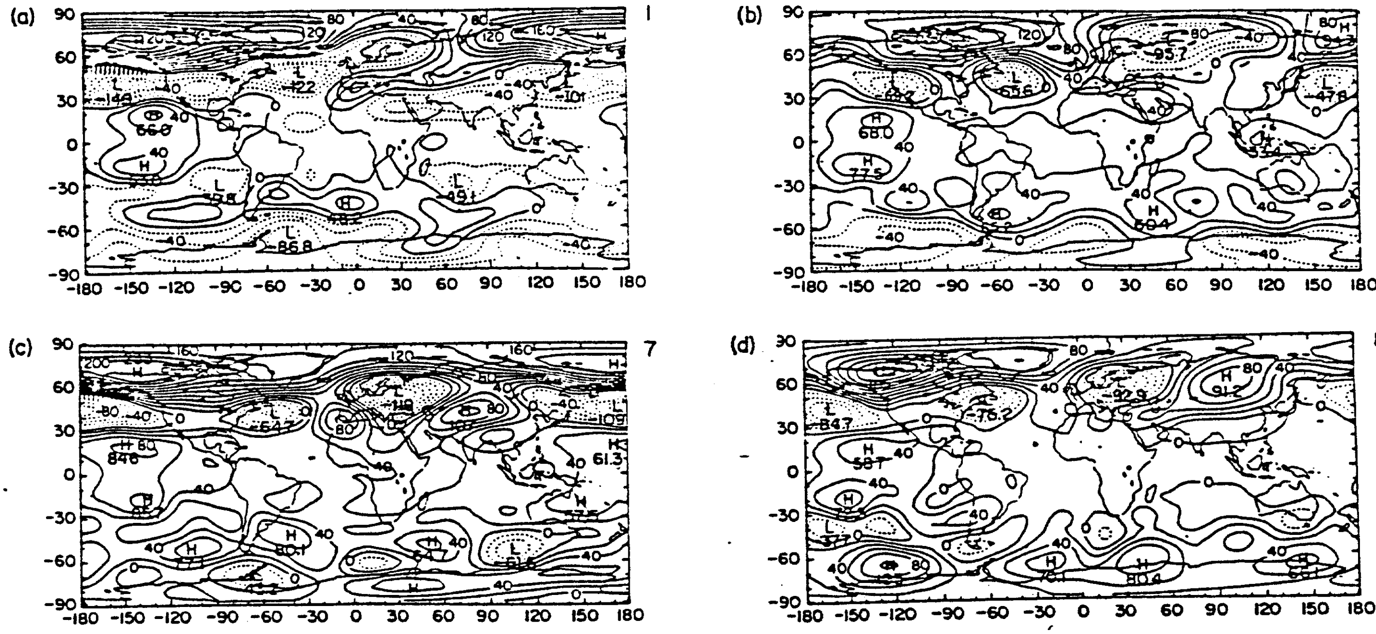

Fig. 5 Four of the eight 90-day average 200 mb height anomaly patterns that go into the ensemble-average anomaly pattern in Fig. 5a.

5. Generation of Midlatitude Patterns by Anomalous Tropical Precipitation

The dominant feature of the ensembled-average simulated 200 mb height anomaly (Fig. 4a) pattern in midlatitudes is a low over the Pacific, a high over Alaska, and a low over Newfoundland. We call this pattern the Pacific/North American (PNA) pattern. It shows up, although rather weakly, as the characteristic pattern of midlatitudes when data is composited for all of the warm equatorial Pacific SSTA events in the years 1948 to 1979 (van Loon and Rogers, 1981). Patterns in individual events vary widely. The 1982-83 pattern (Fig. 6) is at best a highly distorted PNA pattern; the winter of 1976-77, on the other hand, exhibited a classic PNA pattern. In any case, the conventional thinking until very recently has been that the PNA pattern is an atmospheric wave resulting directly from the increase of precipitation (latent heat release) over the warm equatorial Pacific SSTA. The principal question or uncertainty, as reviewed in this paper, has been whether or not this wave is of sufficient amplitude to stand above the noise in midlatitude winter.

An additional complication has been introduced by recent studies (Simmons et al., 1983; Branstator, 1985a, b) of the sensitivity of such a midlatitude wave response to the location of the anomalous tropical precipitation that is doing the forcing. The answer that seems to be emerging from this research is that the PNA pattern can most effectively be forced by a negative precipitation anomaly in the Indonesian sector of the equator in the vicinity of 140°E. This is climatologically the region of maximum precipitation during Northern Hemisphere winter in the zone extending westward along the equator from South America to Africa. What these studies suggest is that a relatively small decrease in precipitation near Indonesia is more effective in forcing the PNA pattern than a much larger increase in precipitation in the central and eastern equatorial

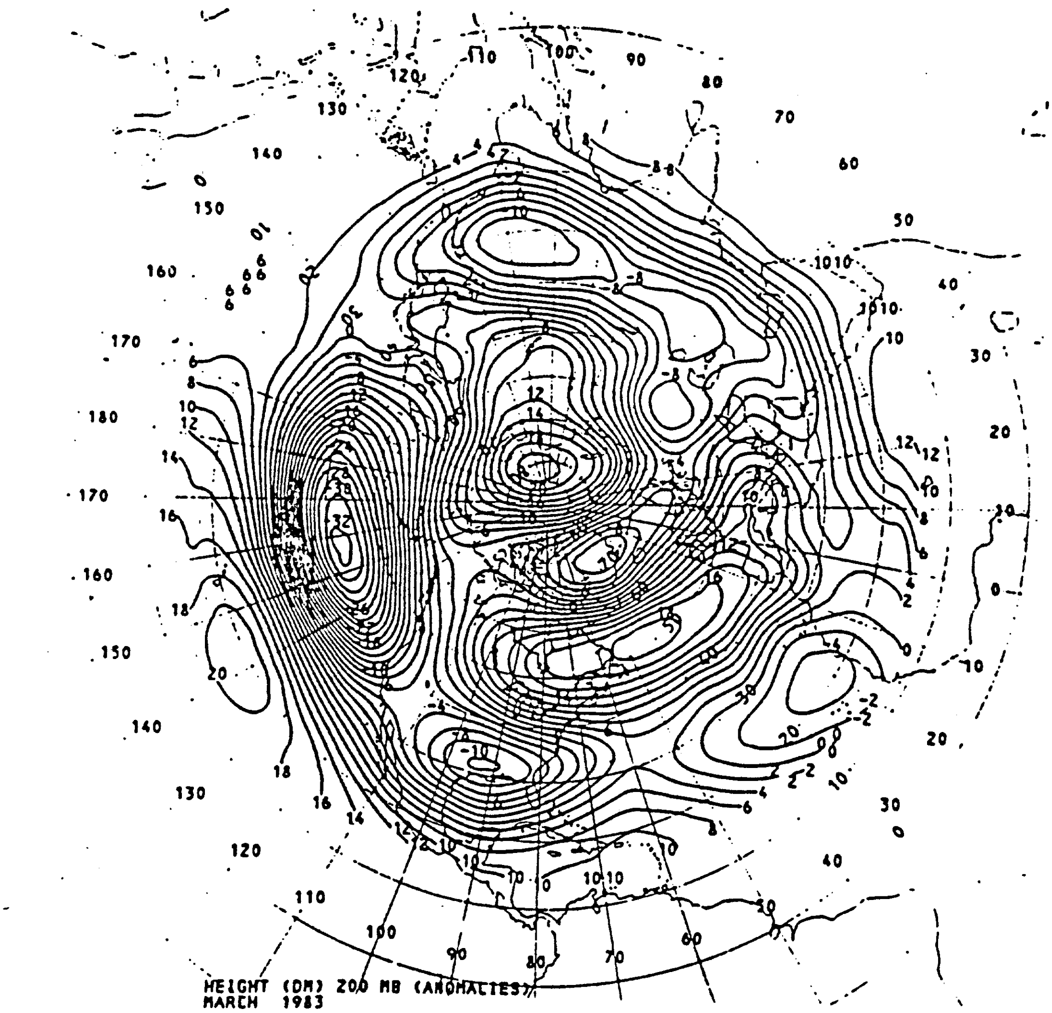

Fig. 6 The observed 200 mb height anomaly in March, 1983; illustrative of the pattern prevailing through the winter of 1982-83.

Pacific. The connection between a warm equatorial Pacific SSTA and the PNA pattern may thus be a much more indirect one than we have thought, proceeding somewhat as follows. The emergence of a warm equatorial SSTA in the central or eastern equatorial Pacific is attended by a positive precipitation anomaly over the SSTA. The primary effect of this may be the forcing of a circulation cell in the equatorial plane with the downward branch of this cell extending westward far enough to suppress the climatological precipitation maximum over Indonesia, this in turn forcing a PNA pattern in middle latitudes. Such an additional link in the causal chain provides an opportunity for yet more variability between individual events in midlatitudes. It even raises the possibility that interannual variability of Indonesian-sector precipitation in years when there is no warm SSTA in the central and eastern Pacific could produce a PNA pattern climate anomaly in midlatitudes.

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SESSION 4

GREAT SALT LAKE AS A HAZARD

Some Perspectives on Climate Hazards: The Case of Great Salt Lake Rise

William E. Riebsame
Natural Hazards Research and
Applications Information Center
University of Colorado
Boulder, CO 80309

Introduction

The fluctuating level of the Great Salt Lake presents a unique set of problems to resource managers in Utah. While the lake is certainly a valuable natural resource, its fluctuation may best be thought of as a natural hazard. This hazard, however, entails several physical and social characteristics that separate it from the typical set of catastrophic events (e.g., earthquakes and flash floods) that are the target of most hazard management strategies. For instance, though lake level rise might conceiveably be_ treated, for policy purposes, as a flood. However, in this case because of lack of a delineated floodplain, and consequent inapplicability of national flood insurance and floodplain management programs, the problem will not quickly be ameliorated by traditional approaches to floods and other natural hazards. Such problems stem partly from the weaknesses of hazards concepts based on a rigid categorization of events, but they also derive partly from the unique problems of managing a climate-sensitive, arid-zone terminal lake. Thus, it may be helpful to describe the problem as a climate hazard.

"Climate hazard" refers to environmental conditions that threaten human property and activities and that derive chiefly from some change or anomaly in a region's temperature and precipitation conditions. Unlike more catastrophic natural hazards (e.g., earthquake, hurricane, or tornado) climate hazards and their related effects usually develop slowly and persist over a relatively long period. This may provide the time necessary for decision-makers to explore a wide range of alternative responses, to compare a full set of costs and benefits associated with different adjustments, and perhaps to formulate and implement innovative, anticipatory adjustments. Yet, like other natural hazards, climate fluctuations involve uncertainty. Will the trend continue (and for how long)? Is the change permanent? These questions are the purview of other papers in this volume. The goal of this paper is to explore the salient characteristics of climate hazards such as lake level rise, and to relate them to Great Salt Lake policy and management.

The Nature of Climate Hazard

The concept of natural hazard (see Burton, Kates, and White, 1978) assumes that humans adjust their activities to some acceptable level of risk from environmental extremes. This equilibrium model recognizes that changes in either social or natural factors can lower or increase the hazardousness of a place, though natural hazards research varies considerably in focus, from mostly human to chiefly physical. The field of hazards research and management, as it has developed in the United States, is fixed chiefly on catastrophic physical events like floods, earthquakes, tornadoes, and hurricanes, with functional emphasis on hazard assessment, human perception and behavior, and hazard mitigation through land use changes and other non-structural adjustments. The field's focus on rather singular, low-probability extreme events does not easily lend itself to a consideration of enduring or recurring hazards like lake level fluctuation, or climate fluctuation in general. However, the growing field of research on climate impacts (see Kates, et al, 1985) offers some insight into the problems of adjusting to climate fluctuation. The relevant concept here is that human activities adapt to a "normal" range of expected climate conditions; extension either above or below this range changes a natural resource into a natural hazard (Heathcote, 1985).

Several characteristics of climate hazards have important policy and planning implications; some of these characteristics differ greatly from those associated with short-duration, catastrophic events (modified from Riebsame, 1985):

Non-catastrophic impacts: unlike catastrophic hazards, climate hazards tend to result in a low-level, cumulative threat to resources (e.g., depletion of energy supplies during a particularly severe winter). Rather than sharp periods of physical destruction requiring emergency rescue, recovery, and clean-up, climate hazards slowly exacerbate resource management problems.

Gradual onset of hazardous conditions: persistent periods of abnormal moisture or temperature conditions tend to result in cumulative effects that are relatively slowly manifest. For example, climate events, such as the 1980 heat wave in the central U.S., bring on slowly accumulating stresses that result in heightened impacts over a long period.

Large areal extent: climate hazards are generally caused by regional or larger scale anomalies in atmospheric conditions. Thus their impacts often range across large areas. For example, the 1980 heat wave covered much of the southern and central U.S. for more than four months, and wet conditions leading to Great Salt Lake rise have affected most of the Mountain and Pacific West.

Persistence: climate anomalies, such as below or above normal precipitation, tend to persist from one season to the next, and perhaps from one year to the next. This persistence was evident, for example, in recent cases of extremely cold winters in the northeastern U.S. (Karl et al., 1984), and in stormy conditions in the western states beginning in 1982 (Rasmusson and Wallace, 1983).

Pervasive rather than intensive impacts: climate flucutations tend to affect several different economic sectors simultaneously, and tend to result in impacts that pervade spatial and social structures. In contrast to concentrated impacts affecting limited areas and population groups, climate impacts result in dispersed impacts that may result in huge aggregate costs.

Climate hazards do exhibit, however, at least one important characteristic in common with other natural hazards:

Uncertainty: like other natural events, future magnitudes and trends of climatic variables are uncertain. Current forecast technology offers only limited skill beyond about 30 days, although some improvements in 90-day forecasts of climate conditions have been achieved. Beyond that, no sound, scientific basis for forecasting currently exists (American Meteorological Society, 1983).

Management Implications of Climate Hazard Characteristics

The characteristics of climate hazards listed above pose a set of unique management challenges and opportunities. The non-catastrophic nature of climate hazards takes the pressure off specialized emergency management services (e.g., search and rescue). Their relatively slow onset offers decision-makers the chance to formulate and implement anticipatory rather than simply reactionary adjustments. However, the "creeping" nature of climate hazards may act against recognition of their seriousness, delaying official declarations of emergency, and hindering the sense that "something out of the ordinary", something requiring special attention, is occurring.

The relatively large areal extent of climate hazards may affect both the availability and demand for hazard resources, and may decrease the ability of jurisdictions to share services. The tendency of climate abnormalities to persist from one season or year to the next requires response and mitigation efforts over a relatively long period, as opposed to the short emergency periods associated with other natural hazards. Finally, because climate hazards often result in pervasive impacts that spread widely through social systems, full impact assessment and mitigation are difficult.

Policy Implications of Uncertainty

Solutions to hazard management problems seem to fall into three broad categories: 1) best fit, 2) maximum flexibility, and 3) least regret. The first approach assumes reliable knowledge of future conditions, for which prescriptive actions, aimed at obtaining the best fit between human activities and the hazard, can be offered. However, most hazard situations involve some uncertainty, suggesting that policy prescriptions calling for a closely engineered solution may be too limited or rigid. Some responses, especially those involving large capital investments and engineering works, tend to limit the range of alternative future adjustments to a problem--if only because managers are loath to abandon projects to which large sums of money have already been devoted.

An alternative to best-fit strategies is to develop policies that attempt to maintain maximum flexibility in responding to future conditions. By maintaining large buffer capacities, or by developing multiple strategies with several options for avoiding future losses, managers can avoid sinking all of their resources and hopes into one, fixed alternative. But such an approach can be quite costly. Indeed, the apparent high cost of multiple, flexible strategies is probably one of the factors leading to the popularity of single-alternative, "best fit" solutions.

One way of modifying best-fit or maximum flexibility strategies is, at the least, by biasing them away from actions that might conceiveably result in increased future losses. Such a "least-regret" strategy might, for example, entail the articulation of worst-case scenarios--what might happen if a certain solution is eventually overwhelmed? If a policy-maker can imagine a logical stream of events leading to enhanced future loss, or even the potential for catastrophic failure of an alternative response, then it makes sense to delete that response from consideration or to provide back-up adjustments.

Finally, within each of the three policy types suggested above, there exist two basic modes for responding to climate hazards such as lake fluctuation: prevention and adaptation (see, for example, Schelling's (1984) review of policy responses to possible climate change due to carbon dioxide). Preventive policies attempt to control hazardous events. Adaptive policies are aimed at adjusting human activities to minimize the losses from hazards.

The Case of Great Salt Lake

For practical, descriptive purposes, one can argue that sometime in 1982 northern Utah experienced a "climate change" (although climatologists agree to no set definition of this term) to wetter conditions that, among other effects, has led to two springs of severe snowmelt floods, landslides and mudslides, and a marked increase in the level of the Great Salt Lake. Although the lake had been rising generally since its historic minimum in 1966 (with some retrenchment during the drought years of the mid-1970s), the rise quickened dramatically in spring, 1983, and has continued to the present. While the attention of local and state authorities was first drawn to snowmelt flooding, it soon became apparent that the floods, and continued wetter than normal conditions associated with below-normal evaporation, had left a legacy of problems associated with inexorable lake level rise, including the threat of inundation to homes, highways, utilities, recreational facilities, and wildlife preserves.

Such marked changes in regional climate, effecting new conditions in hydrological, ecological, and social systems, may evoke large capital investments to protect properties and to maintain activities developed under a different (or what some might have called "normal") set of climatic conditions. In the Great Salt Lake case, lake levels have fluctuated greatly in historic times (Arnow, 1984), but the contemporary lake level problem has its roots chiefly in lakeshore development occurring mostly while the lake level was below about 4200 feet, i.e., c.1930 to c.1973. Thus, for a large part of the current century, land use and development decisions, including the appropriation of new land as the lake receded, assumed either a stable or decreasing lake level. Indeed, at times it was feared that the lake might actually disappear (Deseret News, 1960). Among the developments located on land that had been underwater in historical times were parts of Interstate 80, public beach facilities, the Antelope Island causeway, Willard Bay Wildlife Refuge, sewage treatment plants, a railroad causeway, and several chemical extraction facilities. Since about 1966, with some ebbing in the middle to late 1970s, the lake has risen, already damaging some of these facilities. How long the rise will continue, and how high it will go, are uncertain.

Responding to Lake Rise

Because the recent, rapid rise in lake levels was initiated by a spring of spectacular snowmelt flooding, the creeping hazard of lake rise has now captured the attention of a wide range of governmental hazard management groups. Presidential disaster declarations for flooding and landsliding in 1983 and 1984 brought initial federal attention to the situation; this attention expanded to include the lake level problem (see, for example, Federal Emergency Management

Agency, 1983 and 1984a and b). In some cases, federal funds have been used for lake hazard response (e.g., work to protect Interstate 80), and agencies such as the Army Corps of Engineers are considering possible protective construction efforts in cases where a clear and imminent danger is posed by rising lake waters. Beyond this, no federal disaster assistance is likely to be made avilable for the long-term problem of lake level rise. The chief federal role will be to continue to urge substantive planning to adapt lakeshore land use to future fluctuations (Federal Emergency Management Agency, 1984a).

The responsibility for adjusting to lake level increases remains chiefly with the state and lakeshore jurisdictions (i.e., five counties and several municipalities). This process has actually been going on for several decades. Diking to control certain sections of the lake was first proposed as early as 1861 (see Lake Com Subcommittee, 1973, p. A-71). Serious discussion of major diking, to control lake levels and salinity, started in the 1930s; the chief goal then was to create a freshwater lake east of Antelope Island in order to attract commercial development. More integrated lake control plans appeared in the 1950s. The Great Salt Lake Diking Study proposed several miles of dikes in 1958. The first, preliminary master plan for the lake, developed in 1965, proposed even more diking, with the goal of stabilizing water levels on the southeast shore at 4,200 feet (Great Salt Lake Authority, 1965).

Throughout much of the region's development, concern was focussed on decreasing, not increasing, lake levels. But recurrent lake rises have caused problems throughout the region's history. Clark and Helgren (1980) noted that high levels affected salt harvesting in the late 1800s, and that problems occurred during lake rises following the 1905 and 1933-1940 low spells. One of the first major losses following the 1963 minimum was damage to the Syracuse-Antelope Island causeway built in 1969 on a rising limb of the lake hydrograph. The causeway needed repairs almost immediately (Deseret News, 1969a and b), and was washed out several times before 1973 when it was raised and paved.

Planning that explicitly addressed the problem of lake rise began in a substantive fashion in 1973 when the 40th Utah Legislature authorized a long-range, comprehensive planning effort aimed at managing the lake. Following several studies (e.g., Lake Com Subcommittee, 1973), a comprehensive plan was adopted by the Great Salt Lake Board in 1976 (Great Salt Lake Division and Board, 1976; see also Burnham, 1980). This was followed closely by the Great Salt Lake Resource Management Study in 1977. Record high lake inflows in 1971 and 1975 (Arnow, 1984), and a 50-year record high lake level in 1976, had apparently increased awareness of the need to plan for possible future higher levels. But the dry years from 1976 to

1978 reversed the rise, decreasing fears over possible inundation.

The situation changed markedly in late 1982, with sharp water rises and emergency responses. The Utah legislature allocated $13 million in September, 1984, as part of a flood relief bill. These funds were added to federal and local monies, making available about $44 million for flood damage repairs, mitigation efforts, and perhaps most importantly, research to determine which responses are technically and economically efficient.

The Bureau of Economic and Business Research at the University of Utah conducted a quick study in late 1983 to assess possible losses associated with alternative future lake levels (Bureau of Economic and Business Research, 1983). Such analyses, based on more detailed hydrological probabilities, are continuing at the Utah Water Research Center (James, this volume; see also Bureau of Economic and Business Research, 1984). The goal is to give decision makers an idea of the relative benefits and costs of alternative responses to lake level rise. The net result of such studies seems to be a growing consensus that structural control strategies are economically justifiable in many cases. Development of lakeshore property during periods of low lake levels has placed large capital investment at risk, thus increasing the pressure to choose preventive rather than adaptive strategies. Decision-makers, however, must face the risk of investing in lake control measures that might be unnecessary if the waters do not continue to rise.

Selecting Alternative Responses

Despite the rapid lake rises of 1983-84, it is widely assumed that future lake level changes will typify the less dramatic nature of earlier trends. Slow onset of lake level changes allows for some flexibility in the design of protective works. As James (this volume) points out, control structures must not necessarily be built now for .01 probability lake levels, but can be designed for retrofitting to higher water levels if needed. Slow onset also offers flexibility to non-structural methods, perhaps allowing land uses that are, in essence, "portable". A wide range of alternative response strategies has been proposed, ranging from pumping lake water across a low divide into evaporation ponds on the western desert, to complex transfers of water rights into the Sevier River (Utah Department of Natural Resources and Energy, 1983). A much smaller set of options is being actively considered (see, for example, Allen, et al., 1983).

Most attention is currently focussed on the West Desert Pumping scheme, which, according to a recent feasibility study (Eckhoff, et al., 1983), is the best short-term solution to lake rise (see also Utah Department of Natural Resources and

Energy, 1983). If adopted, the plan would probably be implemented in conjunction with some form of eastern and southern shore diking (Montgomery Consulting Engineers, 1984). This strategy is very compelling because it can, within reasonable limits, literally control the lake surface, and can apparently be implemented without major social or environmental impact. Indeed, the pumping scheme fits the letter of the 1976 comprehensive plan, which requires that lake management maintains a brine concentration in the lake which is economically useful to the chemical extraction industry. The scheme includes a brine return canal which would serve that purpose.

In focussing on the pumping and diking scheme, one could argue that Utah decision-makers are following a path of "best fit" engineering that might limit future options. The unprecedented rise of the lake in 1983 and 1984 has provided the impetus to search for quick-fix, short-term solutions. While it is always difficult to appreciate fully the pressures brought to bear on decision-makers during crises like the current lake level rise, it might be worth pointing out that the great uncertainty surrounding future lake levels probably argues against closely engineered, single-purpose solutions. Indeed, one could argue that, without some form of non-structural change in land use along the lakeshore, the appearance of "control" inculcated by the pumping and diking alternative might actually increase near-shore development and thus heighten the potential for future, perhaps catastrophic, loss if the system fails--i.e., it is not a path of "least regret".

Alternative strategies are on the discussion table even if they are not receiving as much public attention. The Great Salt Lake Contingency Plan (Utah Department of Natural Resources, 1983) offered a wide range of options. It agreed that pumping was probably the best short-term solution, but suggested that decision-makers should also consider longer-term solutions such as further development of up-stream water uses and zoning by local jurisdictions that would designate land below a given elevation (e.g., 4209 feet) for use in agriculture or flexible recreational development. This last suggestion is the equivalent of a maximum flexibility solution, and fulfills the 1976 comprehensive plan's requirement that a "floodplain" be designated around the lake and managed as a "hazard zone". Because this approach has thusfar received the least attention in discussions of how to respond to rising lake levels, the remainder of this paper is devoted to exploring the feasibility of effectively managing a lakeshore hazard zone.

Prospects for Managing a Lakeshore Hazard Zone

The problems prospective to managing a hazard zone around a lake sensitive to climate fluctuation are large and complex.

Fortunately, a great deal of relevant national and local experience has accumulated over the last few decades. For example, coastal zone hazards (focussing on storm-surge and erosion) have been the subject of numerous studies (Office of Ocean and Coastal Resource Management, 1983) and federal and local hazard management schemes. Most appropriate to the Great Salt Lake problem might be efforts by Great Lakes states to deal with fluctuating lake levels and shore erosion: see, for example, Wisconsin Department of Natural Resources (1977), Wisconsin Coastal Zone Management Program (1977), and Great Lakes Basin Commission (1978). The Coastal Zone Management Act incorporates five basic concepts for hazard management, along the U.S. coastline, that apply to the Great Salt Lake case (paraphrased from Office of Coastal Zone Management, 1976):

1) areas potentially affected must be carefully delineated and this information made widely available;

2) estimates of vulnerability must recognize that human occupation of a vulnerable area always involves both benefits and hazards;

3) the full range of possible adjustments to a hazard must be identified. It is rare that only one course of action is worthy of consideration. The full theoretical range of possible adjustments includes:
   - new or improved preparedness plans;
   - control and protection works;
   - design of facilities to resist hazard impacts;
   - land management to minimize loss of life and property, including land acquisition, zoning, subdivision regulations, building ordinances, and easements;
   - insurance against loss; and
   - relief and rehabilitation assistance.

4) an assessment of present and future impacts of adjustments must be made. What is chosen as an adjustment in one place and time may affect the hazard elsewhere or at a later time. Adoption of an adjustment such as beach protection or zoning regulations, unless properly planned, can make matters worse rather than better;

5) there must be recognition that reduction of exposure to frequent events may actually build a potential for catastrophic losses from very rare events. Protective works, for example, may reduce losses from more frequent high water levels while increasing eventual social dislocation from a very rare event.

Platt (1976) identified seven key legal mechanisms by which states have regulated their coastal zones: comprehensive coastal plans, shoreline zoning, state floodplain management, critical area management designations, wetland laws, setback or encroachment lines, and beach or shore preservation laws. A mixture of such strategies would seem to be appropriate in the Great Salt Lake case.

Similarly, floodplain management professionals have, through trial and error, developed a powerful set of strategies for reducing flood losses, especially through non-structural adjustments (e.g., adaptive rather than preventive schemes; for a recent review of non-structural measures in floodplain management see Institute for Water Resources, 1983). Numerous handbooks and guides for implementing regional and local floodplain management are available (e.g., U.S. Water Resources Council, 1981), and the national floodplain management experience has been reviewed in several publications (e.g., Changnon, et al., 1983; see Platt, 1980, for selected case studies). Of special interest in the case of defining and managing a flood hazard zone around Great Salt Lake are studies of legal liability involved in such efforts. Kusler (1983), for example, points out that although communities may be liable for damages due to failure of a flood control system, they are more open to liability if they do not have some form of a "rationally thought-out, combined floodplain management and stormwater management plan" (p. 114). Such a plan establishes a standard of "reasonable care" that makes proof of negligence very difficult. He goes on to note that communities may be especially at legal risk if they are aware of flood hazards but do not formulate reasonable flood management plans.

Experience with both coastal and riverine flooding suggests several guidelines for hazard zone management. For example, during the hazard area delineation procedure, it must be shown that reasonable use has been made of all available scientific information and that the public will have ample opportunity to examine and review hazard designations. It is important to make explicit the full list of alternative adjustments, and it might be useful to prepare scenarios of future impacts, given continued hazard zone development, whenever public regulation of private land use is being considered. The plan must also take into account possible changes in population, technology and risk tolerance that might affect the mixture of desirable adjustments. Any discussion of the adoption of certain adjustments should include recognition of the trade-offs which the community will experience in choosing one option over another, and should outline the way that costs and benefits will be distributed among different groups. In presenting alternative choices, it is also important to identify options which would avoid or preclude irreversible changes or potential larger future impacts (e.g., options associated with a policy of least regret).

Finally, the delineation and management of a hazard zone must include full citizen participation. Effort should be made to evaluate which channels of information have the most credibility in the view of the people for whom the information is designed, and to use those channels to enhance public understanding and involvement.

Research for a Hazard Zone Plan

The foregoing discussion provides some room for optimism in the development of a hazard management zone for the Great Salt Lake. There exists a wealth of experience with similar efforts elsewhere in the U.S. Thus, one step towards such a plan should be a quick evaluation of similar efforts in analogous settings. For example, it may be useful to study Louisiana's experience with wetlands management, especially in the face of sea level rise and land subsidence in the Mississippi River delta. It would also be valuable to review shoreline management plans in the Great Lakes states. Finally, several hazard management programs have included the delineation of special hazard zones, from earthquakes to landslides. This experience should also be surveyed for ideas applicable to the Great Salt Lake case.

A Window of Opportunity

Northern Utah's recent switch to wetter, cooler climatic conditions and the rapid rise of Great Salt Lake are certainly dramatic events, but hazards associated with climate fluctuations have occurred throughout the region's history. As short-term solutions to recent lake level increases are discussed, the opportunity arises to fulfill the requirements of the comprehensive lake management effort started in the early 1970s. Despite climate uncertainty, relatively high lake levels are likely to continue into the near future simply due to the huge influx of water in the last two years. An unfortunate combination of wind speed and direction in some near-future storm could do considerable damage on the south and east shores, turning this creeping climate hazard into a catastrophic event. The raison d'etre for rational hazard zone management around the lake is thus well established. The opportunity to develop logical and socially acceptable management plans is great. But this window of opportunity may not be open for much longer.

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THE USE OF HYDROLOGIC DATA ON THE ECONOMIC AND FINANCIAL ANALYSIS OF LAKE LEVEL CONTROL ALTERNATIVES

L. Douglas James, David S. Bowles, and Eugene K. Israelsen
Utah Water Research Laboratory, Logan, UT

INTRODUCTION

The rising surface of the Great Salt Lake is forcing increasing numbers of property owners to decide between protecting their holdings or moving out. Federal, state, and local governments are considering levees to protect critical facilities, lake level control through pumping to the West Desert, and through water development to increase consumptive use. Both private and public choices are being made with uncertainty as the lake could rise to cause devastating damage or recede harmlessly. Governmental agencies know that many people face economic disaster if they fail to act and the lake continues to rise. Action is expensive; and if the lake falls, the agencies face charges of poor management. We must recognize that while we cannot predict when flooding will occur (through continuation of the present rise or during some future event) we can express risk with a probability distribution, estimate damages in expected values and compare the expected reduction in damages with costs.

STUDY OBJECTIVES

The hydrologic risk quantified in Bowles and James (1985) was used:

1. To examine the economic feasibilities of 19 possible shoreline levees, a levee to enclose Farmington Bay, and West Desert pumping.

2. To estimate residual damages and the consequences for public safety of various levels of protection in order to make recommendations on design levee height.

3. To apply the economic evaluations in refining project designs and reducing project cost.

4. To determine how the benefits are distributed among property owners (public and private) as guidance for cost sharing on individual levees.

PROPERTY MANAGER DECISION SCENARIOS

The damages caused by a lake level rise depend on what the property owners do to protect themselves. Once a protective investment is made, subsequent damage will be reduced by an amount that also depends on the maintenance effort. This opportunity to invest in protective measures as the flood waters are rising distinguishes terminal lakes from other flood hazards. It causes losses to vary with investment and maintenance programs and requires that the damages from rising lake levels be estimated by simulating manager response rather than by using a stage-damage curve.

Managers respond in three phases, self-protection, abandonment, and restoration. As lake levels rise around a property, groundwater and storm waves cause increasing damage. If the benefits of his use justify cost, a property owner will pump to hold down the water table and build levees to keep out surface water. During the self-protection phase, the losses are the costs of required capital investments, operations (largely pumping) and maintenance all of which increase with lake level. Should the lake continue to rise, the cost eventually exceeds the occupancy benefits and the enterprise is "wiped out," and abandons the site.

After wipeout, the loss is the reduction in income to the property owner and continues as long as an alternative site of equal value is not found. As the lake falls, the property owner has increasing incentive to reinstate his facilities to stop further losses. He is not likely to return as soon as the lake level drops below the original site because a reverse could cause another wipeout. However, if the lake falls low enough for long enough, he will accept the risk and reinstate the facility. The cost of reinstatement becomes the third type of loss. Afterwards, the losses have the same nature as in the first phase, but hopefully the amounts are less because the property manager incorporates protective features into his reinstatement design.

DATA SOURCES

Simulation of losses in three response phases requires data for 1) estimating the capital, operating, and maintenance costs of protective investments, 2) quantifying decision rules for abandonment, 3) estimating income losses during abandonment, 4) quantifying decision rules for reinstatement, and 5) estimating reinstatement costs. The hydrologic risk data used were taken from James et al. (1979), James et al. (1984) and Bowles et al. (1985). Levee costs and property values were from preliminary designs and surveys by James M. Montgomery Engineers (1984). The Bureau of Economic and Business Research (1983 and 1984) estimated lake impacts from discussions with property managers. Eckoff, Watson, and Preator Engineering (1983) supplied a preliminary design for the West Desert pumping schemes.

Lake level rises were defined from 100 stochastically generated sequences of lake stages (Bowles and James 1985) that represent distributions of possible future lake levels. These damage sequences were used to determine the mean and the distribution of the damages that might occur. The estimates assume that the recent rise is caused by a random combination of the processes determining weather.

MASTER LIST OF LEVEES AND DAMAGE CENTERS

Nineteen potential shoreline levees were evaluated (Figure 1). Of these, 12 were classified as high priority and 7 as low priority according to a criterion of protecting "public health and safety" specified by Governor Matheson in December 1984 (Table 1). Many types of property are affected by the rising lake, and each type requires different criteria for wipeout and reinstatement and has distinctive loss characteristics during the three phases. For combining information, a damage center was defined as the property of a given type that is protected behind a given levee. A total of 119 damage centers incorporate 14 property types. Table 2 lists the levees and the types of property that they protect.

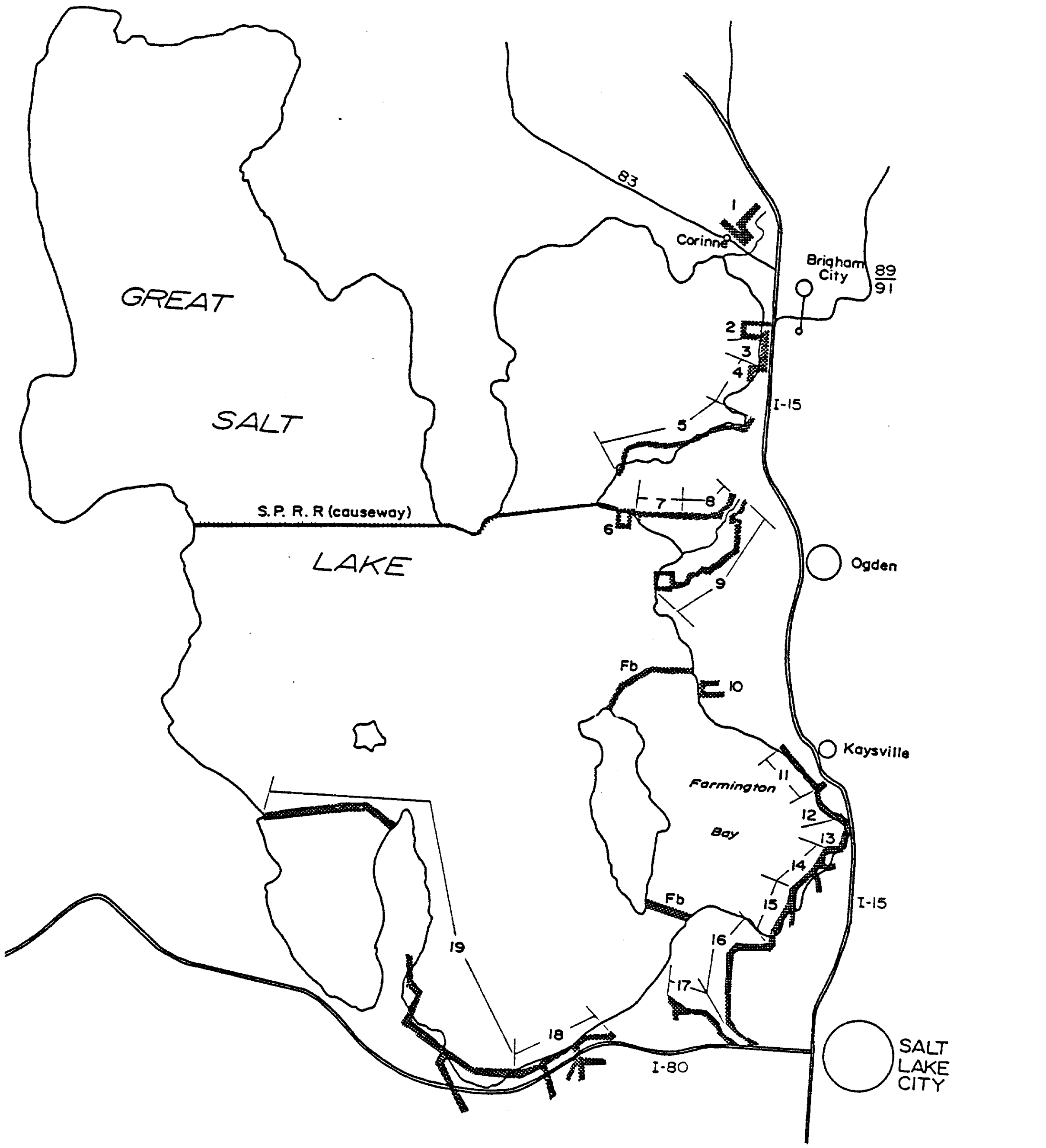

Figure 1. Locations of levees defined on Table 1.

Table 1. Summary of property types by levee.

List of 19 Local Levees

1. Corinne - HP[a]
2. Perry Lagoon-HP
3. Perry-LP
4. Willard Lagoon-LP
5. UPRR-Weber-HP
6. Little Mountain Lagoon-HP
7. Little Mountain-LP
8. West Warren-LP
9. Taylor-Hooper-LP
10. North Davis-HP
11. West Kaysville-HP
12. Wheeler Farm-LP
13. Centerville-HP
14. South Davis Group-HP
15. Jordan River-HP
16. Airport-HP
17. Industrial Park-LP
18. South Shore-HP
19. Tooele County Group-LP

Number of Damage Centers Used to Represent Property Types by Levee

Property Type	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	U[b]	Total
A	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	3	1	5
B	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	2	1	4
C	0	0	0	0	1	0	0	0	0	0	0	0	1	0	0	0	0	0	1	0	3
D	0	0	1	0	0	0	0	0	0	0	0	0	2	0	0	0	0	2	2	0	7
E	2	0	0	0	2	0	0	2	2	0	0	0	2	2	2	0	0	2	3	4	23
F	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	3	3
G	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	4	4
H	2	0	2	0	2	0	0	2	2	0	2	2	2	2	2	0	0	0	2	2	24
I	2	0	2	0	2	0	0	2	2	0	0	0	2	2	2	0	0	0	0	2	18
J	0	0	0	0	1	0	1	0	0	0	0	0	0	2	2	0	2	1	1	0	10
K	1	1	0	1	1	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	5
L	0	0	0	0	0	0	0	0	0	2	1	0	0	1	1	0	0	0	0	0	5
M	0	0	0	0	0	0	0	0	0	0	1	1	1	1	1	0	0	0	1	0	6
N	0	0	0	0	0	0	1	0	0	0	0	0	0	0	0	1	0	0	0	0	2
Total	7	1	5	1	9	1	2	6	6	2	4	3	10	10	10	1	2	7	15	17	119

[a]High priority or low priority as determined in a public meeting according to criteria specified by Governor Matheson.

[b]Property that would not be protected by any shoreline levee.

Table 2. Ranges of parameter values by property type.

	Property Type	IUCS	IUCS1	NVW01	REIN	NSTLOC	NYRIN	IUCSB
A	Mineral Extraction Companies	4210 to 4220	2 to 4	3 to 7	0.5[a]	P	5 to 10	5 to 12
B	Main-Route Railroads[b]	4212 to 4220	2	3	0.15	P	3	8
C	Spur Railroads	4217 to 4220	-	-[c]	0.15	P	5	5
D	Interstate Highways[b,d]	4212	2	3	-	S	-	-
E	Other Roads[e]	4204 to 4212	-	-[c]	0.25 to 0.35	L	2	2
F	Waterfowl Refuges	4212[f]	3	3	-[g]	F,S,P	3	6
G	Recreation Areas	4208 to 4211	3	3[h]	-[g]	S,P	2	2 to 7
H	Agricultural Lands	4210 or 4215	1	2	0.20	P	5	2
I	Residential Buildings[i]	4213 to 4215	2	2	0	P	20	0
J	Industry	4215 or 4220	2	3	0.50	P	5	5
K	Wastewater Treatment Lagoons	4210 to 4212	2	3	0.70	L	3	8
L	Wastewater Treatment Plants[b]	4213	1	3	0.40	L	10	10
M	Utilities	4220	-	-[c]	-	P	-	-
N	Airports	4220	-	-[c]	-	L	-	-

[a]One company was assumed not to reinstate.

[b]Values given are those used to force moving to a higher location; costs are given to raise and maintain other sites as necessary.

[c]Economic wipeout was not used under the assumption that the lines would not be abandoned as long as industry (or agriculture) remained there to be served.

[d]Reinstatement to the lower elevation was not permitted.

[e]Values given are for county roads that would be abandoned with high lake levels.

[f]For facilities remaining operative at the end of 1984.

[g]Reinstatement costs individually estimated by refuge managers.

[h]Economic wipeout was not used for publically owned recreation facilities.

[i]Values given are those that would force residents to move out of floodplain.

MODELING APPROACH

The lake level modeling begins from the initial lake stage, and the damage modeling begins from the initial property development around the lake. Future floodplain development would add benefits. A trace of annual lake level highs is used to estimate damage one year at a time. In the process, an annual high is read. If it is a new high for the trace, wipeout criteria are used to determine whether the property manager would invest in protective measures at a cost of DCI and obligate himself to a maintenance cost of DRM thereafter or suffer a loss DRLI. All three amounts vary with lake level.

Abandonment can be decided because the benefits from site occupancy do not justify either further protective investment should the lake rise higher or continued coping should high levels continue for a long time. For quantifying these two criteria, the highest level for investment in short term protection is specified as IUCS. A long term wipeout occurs when the lake holds within IUCS1 feet of this elevation for NYW01 years. As examples, long term "wipeout" can be caused by reduced salinity in brines available for mineral extraction or the frequent closing of a transportation route by wind waves.

During "wipeout," no further capital investment is made in protective facilities nor are any funds spent for maintenance. Should the lake fall, a manager may determine that he could recoup the investment required to reoccupy his former site before the lake rises again. The simulation reinstates a use at a cost of REIN when the lake remains IUCSB feet below IUCS1 for NYRIN years. Wipeouts can reoccur should the water rise past the wipeout elevation again. Costs of installing and maintaining protective measures, income losses incurred in using or while not being able to use affected property, and reinstatement costs are simulated year by year over the length of the trace. Their totals are summed and also calculated as present worths at a discount rate of 8 percent. Levee costs are estimated and compared with damages prevented. Lake level control schemes, specifically pumping into the western desert, are evaluated by generating a lower sequence of lake levels and summing reductions in the costs of levees and in residual damages.

For initial modeling, the parameters used for deciding wipeout and reinstatement were assigned the values shown for the 14 property types on Table 2. The simulation model could be used to optimize the decision parameters by minimizing the losses by property type, and the optimized parameters could be used to refine the benefit-cost studies.

The cost data that vary with lake level elevation are specified in 3 vectors in 1-foot increments from 4195 through 4220, the maximum simulated lake level.

DCI is the estimated capital cost of protection spent each time the lake rises to a new high unless the property has been abandoned.

DRM is the estimated annual cost spent on such items as the repair of wave damage, keeping protective facilities functional, and pumping drainage water. The costs are taken according to the current lake level before wipeout and after reinstatement and as zero during periods of wipeout.

DRLI is the annual loss, through reduced income or increased expenditures, due to lake levels outside a "normal" range of 4197 to 4202. Losses specified for the wipeout elevation (IUCS) continue after the lake stage drops below the wipeout elevation until reinstatement.

PARAMETER ESTIMATION

Many factors must be considered in estimating how costs and losses vary with fluctuating lake levels. For example, DRM estimates the increase in maintenance cost as the lake rises. Data were found on costs of 1) preserving the levee crest elevation as a heavier embankment becomes more prone to sink into the foundation muds, and 2) maintenance as water rises closer to the levee top. A higher levee adds weight to increase foundation settlement, and higher water on a given levee adds wave damage. The cost of embankment maintenance also provides for levee patrol. Salt company and railroad data showed a cost increase proportional to the 1.7 power of the lake stage.

DRLI covers losses when traffic is stopped on railroads and highways by waves during wind storms. Routes at the 100-year wave height were assumed to be closed an average of one day every 100 years. At the south tip of the lake, the reach most exposed to wind, the 100-year wave height is 9.9 feet. The wave height on the second day of a storm is about half that on the peak day (Lin 1976), and that decay rate was assumed to continue. Secondly, it was assumed that the second fastest wind day in a year would have a peak velocity about 90 percent of the maximum. Wave heights were taken as proportional to the square of the wind velocity. These day-to-day and storm-to-storm decay rates were used to estimate the number of days of closure for events of different return periods. These were plotted on probability paper to determine the average annual days of closure for various roadway heights above the lake. The results at the south tip of the lake were -0.01 days of closure per year for a route 10 feet above the lake; +9 - 0.03, +8 - 0.07, +7 - 0.12, +6 - 0.45, +5 - 0.91, +4 - 1.4, +3 - 2.5, +2 - 5.0, +1 - 15.0. These durations are 1) multiplied by a loss per day of closure, and 2) adjusted downward for lakeshore reaches where the waves are lower.

Other considerations used in parameter estimation are:

1. Mineral Extraction Companies. The wipeout elevations were judged from management interviews and varied with the financial strength of the company. Reinstatement elevations varied with the profitability of the site. The protective investments was assumed to be spent on dikes. The losses because of brine dilution were assumed to equal half of the annual return on investment at the wipeout stage and vary linearly down to zero at 4205.

2. Main-Route Railroads. The two main railroad routes are the Southern Pacific Causeway across the lake and the Union Pacific Line along the south shore. The SPRR causeway costs were extrapolated from data obtained from the railroad to a maximum lake level of 4220. The UPRR is divided into two damage centers. The lower represents the present alignment, and the wipeout elevation is the maximum reasonable stage for this route. The second represents a relocation to a higher alignment that would replace a 10.7-mile with a 16.4-mile long section. Reinstatement would be a return to the present alignment to reduce shipping cost on a shorter route with a flatter grade. Construction on the new alignment was assumed to take 3 years. The loss for a day's delay in freight transport during wind-wave closures was taken as $10,000 for both railroads. The cost of freight transport on the longer relocated route was estimated

at 2 cents per additional ton-mile for an average annual traffic of 100,000,000 tons per year to give a $11,400,000 annual loss.

3. Spur Railroads. The three spur routes are the UPRR line between Ogden and Little Mountain, the Denver and Rio Grande Line near Centerville, and a UPRR spur in Tooele County. Data on replacement costs, compiled by Montgomery, were used to estimate the costs of raising them. The durations of wave-caused traffic interruptions were based on base elevations midway between the roadbeds before and after raising. The loss was taken at $500 daily.

4. Interstate Highways. The threatened interstate highways are I-15 near Perry, I-80 in Salt Lake County, I-80 in Tooele County, and I-15 near Centerville. The freeway sections where part of the present road elevation is below 4212 were subdivided into two damage centers because the values of DRM and DRLI are large and are reduced by raising or by relocating (as proposed at the south tip of the lake) the freeway. Wipeout of the lower center activates the higher one. Reinstatement is not used. Replacement costs were used in estimating DCI (spread over 3 years) and DRM. Traffic interruption losses were based on $1 per vehicle delayed ($10,500 daily on I-80, $25,000 daily for I-15 at Centerville, and $15,000 daily for I-15 at Perry). The extra travel cost (including the value of vehicle-occupants time) on a longer route should the highway be moved is estimated at 32 cents per vehicle mile. For 3,840,000 vehicles annually traveling 2.6 more miles. The loss is $32 million each year after relocation.

5. Other Roads. The county roads were divided into seven cases defined in Table 3. The wipeout elevation was set at the lowest road elevation found on the contour map for cases 1 and 2. For cases 3 and 4, costs were estimated for raising up to three times should rising lake levels require it, once to 4212, a second time to 4217, and a third time to 4220. After the lake recedes, wave damages are based on the raised elevation. The losses from traffic interruptions were based on $50 per day for local roads and $1000 daily for through roads.

6. Waterfowl Refuges. The federal bird refuges, state bird refuges, and private hunting facilities cannot be protected by shoreline levees, but some would be enclosed within the Farmington Bay levee (Figure 1). The facilities

Table 3. Cases used in simulating damages to county roads.

Case 1:	Roads below 4212 that are not to be raised because they only service local property that will be flooded.
Case 2:	Roads between 4212 and 4217 that are not to be raised.
Case 3:	Through roads below 4212 that would be raised.
Case 4:	Through roads between 4212 and 4217 that would be raised.
Case 5:	The Bear River Bird Refuge Road.
Case 6:	The Stansbury Island Road.
Case 7:	The North Antelope Island Causeway.

are already largely inundated, but 4212 was used for wipeout for the remaining capital facilities. The University of Utah survey gives damage estimates that were assumed to equal the capital costs for protection. The Division of Wildlife Resources estimates 120,000 hunters annually at the lake. An estimated economic value per hunter-day translates into an economic value of marsh land for bird food (James et al. 1979) of about $10 per flooded acre annually.

7. Recreation Areas. The four recreation damage centers (South Shore Park, the Saltair Marina, the Saltair Resort, and the Antelope Island Park) would only be protected by the Farmington Bay levee. Wipeout elevations were determined from the reported interviews with the park managers. James et al. (1979) estimated economic values of $1,105,000 annually from recreation on Antelope Island and $4,170,000 annually at the south shore parks. Visitation during the last few years has been about 35 percent of the levels projected in 1979. Recreation use was assumed to be unaffected by lake levels to 4205 and be zero with lake levels above 4212.

8. Agricultural Lands. Montgomery Engineers (1984) used tax assessment records to sum the market values of agricultural land in two elevation zones. Each zone combines an area of inundation with higher ground affected by high water tables and storm waves. The "value" of the lower zone combined the market values below 4212 with half the market value between 4212 and 4217. The upper zone contained the remaining value of the agricultural land below 4217 plus half the value of land between 4217 and 4222. The two zones were used to define two damage centers even though one could subdivide more finely in a situation where fields vary in elevation. Reinstatement was assumed to take 5 years for salt to be leached from the soil. It was assumed that incomes from farming near the lake are too low to pay for levees. The income lost when fields are abandoned was estimated from an 8 percent return on investment.

9. Residential Buildings. Assumptions were made that 1) residences will be flood proofed against a wave height of 3 feet and will be abandoned as the lake approaches their floor elevation, 2) the cost of flood proofing buildings against 3-foot lake waves equals the cost of flood proofing against 3 feet of riverine flooding (James and Lee 1971, p. 244) or 0.137 times the market value, 3) abandonment will be followed by zoning that will prevent return after the waters recede, 4) people who abandon their residences will rent an alternate home at an annual cost of about 10 percent of the market value, and 5) the rent will be counted as damage for the remaining mortgage period for present ownership. Two damage centers are used behind a levee. The lower one covers the cost of flood proofing, and the higher one covers the cost of moving to a flood-free location. An artifact of zero-cost reinstatement is used to stop simulating a continuing loss for alternative housing arrangements. The cost of vacating and moving was assumed to be 5 percent of the value of the buildings.

10. Industry. Industrial buildings are found in the Little Mountain Industrial Park, South Davis County, Rose Park, the Airport Industrial Park, Chevron, and a commercial area in Tooele County. The principles used to simulate damage followed the rules used for residences except that 1) reinstatement was permitted, and 2) the industrial areas were divided into two elevation bands (as opposed to using damage centers distinguished by before and after moving). Much of the equipment and many of the buildings could be moved to another site and returned or substituted.

11. Wastewater Treatment Lagoons. Sewage treatment lagoons are used to treat small volumes of wastewater on low cost land at Corinne, Perry, Willard, Plain City, and Little Mountain. They would not function properly with salt water intrusion, and that would be costly to prevent in a leveed enclosure. A popular alternative for a small community without access to inexpensive land is package treatment. A plant can be readily obtained, but one is expensive and generally requires a full-time operator. Wipeout was defined as a switch to a package plant when protection of the lagoon becomes too expensive. Wipeout elevations were read for each lagoon from contour maps. The rental cost of a package plant is estimated to be 15 percent of the lagoon cost plus $40,000 annually for hiring an operator. A community could sell the plant should the lake go down.

12. Wastewater Treatment Plants. The threatened physical-chemical wastewater treatment plants are North Davis, Central Davis, South Davis North, South Davis South, and Salt Lake City. Functioning facilities would have to be preserved to protect the lake environment from untreated wastes. The alternatives are either to protect the plant or move it. For the isolated plant at North Davis, the benefits were estimated from the cost savings of a special levee compared to moving. For the other plants, the savings was from a general levee protecting a mix of properties compared to an individual levee. Equipment could be moved to another site, but the full cost was used for relocation to compensate for higher land costs away from the lake.

13. Utilities. The principal threatened utilities are high tension power lines built on trusses 30 to 40 feet above the ground. Some telephone and water lines, a radio tower, etc. are also affected. It was assumed that the towers can be protected to 4220 against ice forces by added concrete in the foundation and added steel in the lower parts of towers. The estimated cost reached 61 percent of the full construction cost at a water depth of 9 feet and suggests replacement by towers specially designed to stand in water. The annual maintenance cost was increased by 50 percent when the line is in the water because of erosion at the base of the towers, the corrosion of steel members by salt water, and greater difficulty in effecting repairs. Losses from power outages were not estimated.

14. Airports. The Little Mountain Air Base on the southern edge of Bear River Bay has already lost its helicopter pads and is not subject to further damage. The Salt Lake International Airport can be protected by pumping plants to drain the airport subgrade and levees between the airport and the lake. Cost estimates were obtained from airport authorities. The amount of seepage water to be pumped increases approximately with the square of the head, and pumping power requirements increase directly with head.

SUMMARY AND ANALYSIS OF HYDROLOGIC DATA

The effects of pumping on selected high sequences of lake stages are shown in Table 4. The probabilities of failure by overtopping estimated from the generated sequences for levees of various heights during a design life to 2050 are summarized in Table 5. Two sequences required pumping for many years, suggesting that the pumping plant be designed accordingly. The effectiveness of pumping in reducing the peak of the current rise is determined primarily by how soon the peak is reached and secondarily by the volume of water during the largest inflow years. The peaks of later rises (one starting from below

Table 4. Effects of pumping on lake levels.

	Sequence 41 with	Sequence 41 without	Sequence 45 with	Sequence 45 without	Sequence 58 with	Sequence 58 without	Sequence 91 with	Sequence 91 without
1984	9.20	9.20	9.20	9.20	9.20	9.20	9.20	9.20
1985	10.38	10.38	12.02	12.02	12.27	12.27	12.12	12.12
1986	10.04	10.83	15.54	16.07	14.14	14.68	14.88	15.42
1987	9.89	11.60	17.20*	18.19*	15.14	16.29	17.27	18.25
1988	10.43	12.81	15.70	17.29	15.89*	17.45*	18.24	19.52
1990	9.12	12.83	8.83	11.75	12.95	15.15	19.00*	20.76*
1992	7.96	12.65	3.17	7.32	9.42	12.82	16.22	18.61
1994	9.97	14.34	1.23	4.38	5.87	10.59	11.29	14.26
1996	12.82*	16.77*	1.82	5.21	5.74	11.06	6.29	10.89
2000	8.25	13.36	-	1.77	6.50	12.59	1.87	7.78
2005	0.82	5.93			1.41	7.69	-	-
Drop in Peak[a]		3.95		0.99		2.56		1.76

[a]Comparative drops for the four highest future sequences are 4.93, 4.39, 2.87, and 4.86.

*Peak years.

Notes: All results are with a pumping trigger at 4202.
All elevations are above a datum of 4200 ft. msl.

Table 5. Probabilities of levee failure before 2050.

Levee Top	No pumping	With pumping: During Cur. Seq.	With pumping: During Subs. Seq.
4212	0.57	0.30	0.06
4214	0.28	0.12	0.03
4217	0.07	0.02	0.01

4202 after the pumping plant is installed) are reduced more. The largest reductions in lake level occur during drawdown with the maximum simulated amount being 7.8 feet. An operating rule for stopping pumping at a higher lake level during drawdown should be examined.

SUMMARY OF THE BENEFIT ANALYSIS

The 20 levee sites were individually evaluated through estimates of the:

1. Present worths, discounted at 8 percent, of the construction, operation, and maintenance costs for levees designed to withstand lake levels of 4212, 4214, and 4217.

2. Present worths of the benefits of levees of each height for protection against the planning scenario of lake rises.

3. Present worths of the benefits of levees of each height and averaged over the 100 stochastically generated scenarios.

4. Comparison of the best estimates of the costs and benefits as an index of the economic efficiency of the levee.

5. Probability distribution of the benefit-cost ratio as information on the range of future conditions in which the levee is justified.

6. Distribution of benefits by property type as information for determining how the costs should be shared.

7. Portion of the total damages prevented (damage reduction efficiency) as information for safety assessment.

Example results are given in Table 6. The levee cost estimates give expenditures by year and the present worth of the cost of providing levee protection over an average expected design life that increases with levee height. The low benefit-cost ratios show that this levee cannot be justified economically. The damage reduction efficiency is defined as the fraction of the total damages prevented by the levees. A high efficiency means that very little damage would continue after the levee is built. A low efficiency suggests caution because levee construction may be a prelude to a disastrous failure.

Table 7 sums the losses estimated for all 119 damage centers. If no control measures are taken, approximately $3.6 billion in damages ($1.2 billion discounted to present worth) can be expected by 2050. Expected losses are much greater in the early years because of present high lake levels. As a discounted average, only about half (587 of 1210) of the expected damage could be prevented by levees. An undiscounted summation shows that only 22 percent (801 of 3638) could be prevented because much of the loss occurs during repeated rises outside the protected area. However, 73 percent (2151 of 2951) could be prevented during a severe rise. The percentage of the benefits accruing to property belonging to state and local governments averages 37 (217 of 587), suggesting the private property owners and the federal government should be willing to pay close to two thirds of the cost of protective measures lakewide. The best estimate of damage to protectable property in 1985 is about $350 million if no protective measures are taken. This could be reduced to about $60 million by levees.

INITIAL ASSESSMENT

The following tentative recommendations come from a longer report (James et al. 1985) based on preliminary data and illustrate issues that can be resolved by combining hydrologic and economic analyses.

A. Design Standard for Levee Construction

Two events are commonly used in the design of dams, levees, and similar hydraulic structures. The smaller event is to be controlled by the structure, and the larger one is used to check the design for safety against disastrous failure. Levee failure because of a severe windstorm or a foundation settlement (possibly associated with liquefaction during an earthquake) could cause billions

Table 6. Example results for the Corinne Levee.

All monetary amounts in thousands of dollars

Property Protected: Total Values: To 4212--684 To 4217--1734
Fractions of Value Totals
County Roads - 0.26, Farm Lands - 0.42; Waterwater Tr. - 0.12; Buildings - 0.15

Levee Cost Estimates

	Construction	Op.	Main	Sum		Design Life
Levee to 4212 (31-year average life)						
Amount	$2425 for 1 yr	3	99	102 $/yr		
Present Worth (8%,65y)	2425	37	1229	1266	3691	3583
Levee to 4214 (50-year average life)						
Amount	$2881 for 2 yr	3	109	112 $/yr		
Present Worth (8%,65y)	2847	37	1353	1390	4237	4217
Levee to 4217 (61-year average life)						
Amount	$3585 for 2 yr	3	124	127 $/yr		
Present Worth (8%,65y)	3499	37	1539	1576	5075	5073

Economic Analysis with Benefits Estimated from Damage Simulations with:

A. Average Results from 100 Stochastically Generated Sequences

	Federal-Private	State-Local	Total
Sum of Damages			
Present Worth @ 0%	520	2642	3162
Present Worth @ 8%	227	991	1218
With Levee to 4212 (31-year average life)			
PW Damages @ 8%	157	491	648
PW Benefits @ 8%	70	500	570
Fraction of Benefit	0.12	0.88	
Benefit-Cost Ratio			0.16
With Levee to 4214 (50-year average life)			
PW Damages @ 8%	87	205	292
PW Benefits @ 8%	140	786	926
Fraction of Benefit	0.15	0.85	
Benefit-Cost Ratio			0.22
With Levee to 4217 (61-year average life)			
PW Damages @ 8%	29	58	87
PW Benefits @ 8%	198	933	1131
Fraction of Benefit	0.18	0.82	
Benefit-Cost Ratio			0.22

B. Benefit Distribution Defined by the 100 Stochastically Generated Sequences Sorted by Benefits

1. Benefit-Cost Ratios for Sequences of the Following Probabilities:

	0.01	0.05	0.10	0.25	0.50	0.90
Levee Top Elev.						
4212	0.41	0.38	0.32	0.31	0.06	0.01
4214	0.41	0.36	0.34	0.31	0.26	0.03
4217	0.39	0.36	0.35	0.29	0.25	0.09

2. Damage Reduction Efficiencies for the Same Probabilities:

PW Damages-$million	2.08	1.83	1.79	1.51	1.28	0.69
Levee Top Elev.						
4212	0.70	0.74	0.69	0.73	0.17	0.07
4214	0.83	0.84	0.79	0.85	0.86	0.20
4217	0.95	1.00	0.98	0.97	0.97	0.66

Table 7. Total damages in millions of dollars.

	Unprotectable Centers S & L	Unprotectable Centers Total	Protectable Centers S & L	Protectable Centers Total	All Centers S & L	All Centers Total
By year:						
1985	5	57	127	298	132	355
1990	5	44	8	25	13	69
1995	5	45	4	20	9	65
2000	4	43	3	15	7	58
2010	4	42	2	6	6	48
2050	4	42	2	3	6	45
Sum	275	2835	301	801	576	3638
PW (8%) mean	63	623	217	587	280	1210
By probability:						
PW (8%) 1% trace*	86	800	629	2151	715	2951
PW (8%) 5% trace	93	748	387	1471	480	2219
PW (8%) 10% trace	86	734	350	1271	436	2005
PW (8%) 25% trace	70	676	310	994	380	1670
PW (8%) 50% trace	63	631	191	481	254	1114
PW (8%) 90% trace	55	570	59	-216	114	354

*The trace percentiles are sorted by benefits with the 1% trace being the largest present worth from any trace.

of dollars in property damage and loss of life. The control standard could be to contain a lake level having a 1 percent chance of occurring in any given year, approximately 4217. The safety standard should be set higher, perhaps at a level associated with 1 percent chance of failure during the entire design life of the project, approximately 4221 or 4218 if pumping is used. The levees need not be initially constructed to this elevation because the rate of lake rise provides ample opportunity to raise the levees later. However, site selection, foundation preparation, and initial construction should permit subsequent raising when needed. Foundation conditions should be carefully examined to make sure that the underlying soil can hold levees the height and weight required by the safety standard.

B. Shoreline Levee Assessments

1. Corinne: The levee is not justified economically; the least cost approach would be to abandon or move buildings out of the inundated area, cease farming waterlogged fields, abandon roads when water forces their closure, and replace the sewage treatment lagoon with a package treatment plant. However, the cost of a package treatment plant would impose a major financial burden on Corinne unless arrangements are made to help the small community with financing.

2. Perry Wastewater Treatment Lagoon: A levee to protect this exposed lagoon is not justified economically. The lagoon performance is particularly in question because salt water would impair the biological activity that provides the treatment.

3. Perry: The levee is not presently justified economically. Contingency planning should continue because a rising lake would threaten the freeway.

4. Willard Lagoon: This levee is nearly justified economically because it is easier to protect at its higher elevation. The lagoon should be operated until it is seriously threatened.

5. UPRR-Weber: The particular levee configuration examined is not justified economically, but its benefit-cost ratio is close enough to unity to suggest refining the design after careful delineation of the topography, foundation conditions, and the locations of property subject to damage. The results should be presented to the owners of the protected industrial property. The decision would depend on whether they would be willing to pay the approximately 85 percent of the cost that would be their fair share.

6. Little Mountain Lagoon: A levee to protect this lagoon is not justified.

7. Little Mountain: Should the lake level rise past 4214, the benefit-cost ratio will be close enough to unity to refine the design for presentation to the property owners.

8. West Warren: This system of levees has an overall benefit-cost ratio close to unity and largely protects residential property. The levees were not given a high priority, but the magnitude of the threatened economic loss and its widespread distribution to small property owners justifies more careful analysis. Homeowners have fewer resources for raising capital than do industrial property owners. An effort should be made to help them raise the necessary money, perhaps through assessment bonds.

9. Taylor-Hooper: This unit protects farmland, scattered residential properties, and an office of the Division of Wildlife Resources, and the benefit-cost ratio is far below unity. Zoning to prevent future residential occupancy would be desirable.

10. North Davis: The protective levee should be constructed at a schedule that will ensure that effective treatment will continue.

11. West Kaysville: A local levee should be constructed to protect the Central Davis STP, but additional levees are not justified economically. UP&L should determine whether it is less costly to reinforce or move its power poles.

12. Wheeler Farm: The full levee is not justified economically, but local protection for high damage areas should be examined.

13. Centerville: The costly Centerville levee protects some very valuable property and has a benefit-cost ratio of close to unity. A cost effective design should be presented to the private property owners.

14. South Davis Group: Most of the property value protected by this expensive group of levees is at the North Davis South STP; this facility should be protected. A few industrial properties should be examined for local protection, and a least-cost method should be sought for maintaining the power lines.

15. Jordan River: This levee system has a very high benefit-cost ratio. The design goal should be to find the least-cost combination of levees, riverbank dikes, and pumps that provide satisfactory protection. A plan should be developed to recover a fair share of the cost (about 90 percent) from the private property owners.

16. Airport: The search for the optimal method of protection should examine foundation conditions, topography, the division of water flowing under the airport between lake and land side sources, alternate pumping configurations, and protection from seiches and high waves.

17. Industrial Park: The levee is not presently justified economically, but the situation should be watched. Future needs should be presented to the property owners.

18. South Shore: The South Shore levee is costly but protects some very valuable property and has a benefit-cost ratio of close to unity. The design should proceed and seek the least cost alignment should the levee eventually have to be raised to 4221 (4218 with pumping). This levee is particularly exposed to windstorms, and special care should be given to wave protection.

19. Tooele County: The strong economic justification for this full levee system demands that construction and cost sharing arrangements proceed. Past experiences of the industries and transportation companies have fairly well fixed the design and siting for the levees.

C. Enclosure of Farmington Bay

The enclosure of Farmington Bay with levees connecting Antelope Island to the mainland at its north and south ends (Figure 1) has a benefit-cost ratio below 0.1 because the shoreline levees provide nearly as much protection for much less cost. The benefits are enhanced by greater recreation opportunities on Antelope Island and by the transportation benefits of having a highway from Ogden to Tooele, but the amounts are small relative to costs.

D. Pumping to the West Desert

Pumping to the West Desert has benefits that far exceed its cost whether the lake continues to rise or stabilizes at about its present elevation. For example, pumping will save about 3 feet of required levee height. The potential savings in levee cost would be considerably more than the cost of pumping. The benefits to private property owners are about four times their fair share of the cost, and it should be possible to devise a way for the state to fund construction and recoup its costs through a tax over time.

CONCLUSIONS

Damage simulation provides a methodology for evaluating the economic justifications of lake level control and levee alternatives and provides guidance (that can be refined as later studies are completed) for determining where more cost effective designs can be found, what should be done to protect public safety, and how to cost share. Further studies should explore refined designs for public levees, local protection for private property, and environmental impacts. The protective system should be integrated with land use policy near the lake. One benefit from shoreline levees would be clear definition of the boundary between the lake and developable lands in urban areas as has occurred with urban growth in every maior city around the world.

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"Federal Hazard Mitigation and the Great Salt Lake"

Clancy Philipsborn
Region VIII
March 22, 1985

> **[As we know] "The Great Salt Lake lies at the bottom of a closed basin. Due to the wide range of inflow to the lake, the surface level, surface area and volume of the lake has experienced wide fluctuations in the recent past. Efforts have been made to predict future levels of the fluctuations to avoid problems of development around the lake that would be damaged by high lake levels. Recent studies have predicted levels to elevation 4212 in the near future. The general consensus of researchers and climatologists is that such predictions cannot yet be made with any degree of assurance. The data should, however, serve as a warning that the lake could rise to levels that would cause considerable damage to new and existing development around the lake."**

These are the exact words of Lloyd Austin, Division of Water Resources, Utah Department of Natural Resources in an abstract to his paper, "Lake Level Predictions of the Great Salt Lake." Written in 1979, his words provide an indication of how well his advice has been heeded. Here we are, six years later, saying the same things.

My name is Clancy Philipsborn, and I work for the Disaster Assistance Programs Division of the Federal Emergency Management Agency (FEMA) in Denver. For the past two years, I have served as the Federal Hazard Mitigation Coordinator in response to the 1983 and 1984 Presidential Declarations of Major Disaster Areas in Utah. It is my job to lessen Federal expenditures from future flooding through 1) the coordination of nonstructural recommendations generated from all involved Federal agencies through an Interagency Team, and 2) the development of a strong State Hazard Mitigation Program that addresses shortfalls in Utah's capability to respond, recover, and prepare for hazardous events through long-term planning principles.

When we speak of mitigation, we mean the reduction or alleviation of the hazard. This can be accomplished by:

1. Eliminating the hazard. An example would be the construction of a dam, followed by the regulation of the downstream flow through controlled releases;

2. Reducing the impact and damage that the hazard is likely to cause. Dikes can keep floodwaters away from investments, and building codes can strengthen designs to withstand such impacts as debris flows, floods, and earthquakes; or

3. Eliminate the impact and damage that might occur. Decisions to not develop new structures or relocate existing ones eliminate potential damage altogether.

Constance Steffans, in her report "Flooding and Landslides in Utah: An Economic Impact Analysis" (1984) has identified, foot by foot, what damages and impacts can be expected as the Great Salt Lake continues to rise. This conference has offered a great deal in assigning probabilities to those incremental events occurring. We must now present this information to the public and the policy/decision makers, so that an acceptable/nonacceptable degree of risk may be determined.

* * *

The Federal government formally adopted a non-structural approach as part of their overall flood recovery strategy in 1980 when analysis revealed that $10 billion had been spent for flood control structures in the last 40 years. At the same time, average annual flood losses for the United States are $3.8 billion per year and rising. Obviously, we will never be able to stop floods from occurring, but it is also apparent that structural solutions alone are not reducing disaster recovery costs. The Federal government has an inherent responsibility to reduce its cost, wherever possible, and our efforts in this field can be easily recognized in many areas. For example, FEMA's disaster assistance programs are much more restrictive than was the case in earlier years. As many Utahns are well aware, we cannot make governments or individuals "whole" again, as in the days of huge block grants. Assistance to State and local governments to repair or replace damaged public facilities is no longer 100% Federally funded, it is a 75%-25% Federal-State cost-share. Eligibility requirements for the temporary housing of individuals are more restrictive. Flood insurance premiums have risen, while coverage has been reduced. Outright grants to cover unmet serious needs are limited to $5,000. Within other Federal agencies, changes in regulations have had their effect too. The IRS has reduced its casualty loss allowance, Small Business Administration and U.S. Department of Agriculture loan programs are targets for possible elimination, and as mentioned, large Federal grant programs from Housing and Urban Development, Community Services Administration, and Economic Development Administration have been significantly reduced or eliminated altogether.

When one adds to these considerations the thrust and implications of recent litigative history, it is no wonder that State and local government's interest and participation in mitigation program planning have increased dramatically. State and local governments are increasingly the targets of legal actions by both individuals and the Federal government for negligence and liability with respect to decisions or lack of action that have led to, or compounded the effects of, disasters.

We often refer to "the four phases of emergency management;" Preparedness, Response, Recovery, and Mitigation. It is essential to realize that whenever a mitigation opportunity is ignored, then there is an equal and opposite preparedness, response and recovery responsibility that must be addressed.

* * *

Federal involvement in mitigation projects is neither new nor limited to FEMA. It is only in recent years that FEMA has begun to develop a coordination and leadership role in hazard mitigation activities. Federal activities can be traced back to the 1930's when land treatment, reforestation, and retention facility projects were undertaken by the Forest Service and through WPA (Works Progress Administration), primarily along the Wasatch Front in Davis County. The National Weather Service has provided meteorological forecasts beginning in the same time period. The Corps of Engineers enlarged the Surplus Canal on the Jordan River in 1960.

Currently, the Federal family is involved in a number of mitigative projects in Utah, many of which directly affect the Great Salt Lake. The U.S. Army Corps of Engineers has had a very visible role. On Utah Lake the Corps diked the Provo Airport and the Lakeview Subdivision just to the southeast in 1982. The Corps would also have dredged the Jordan River last spring, had Utah and Salt Lake Counties been able to reach a timely assurance agreement. The Corps is presently involved in protecting two sewage treatment plants on the eastern shore of the Great Salt Lake, one in Davis County, and one in Box Elder County. The Corps has also provided technical assistance through on-site personnel over the past 2 1/2 years.

The Federal Highway Administration has provided funding, through the Utah Department of Transportation, to protect and raise portions of Interstate 15 by Provo, and Interstate 80 along the southern shore of the Great Salt Lake. The Department of Agriculture has been involved through the Soil Conservation Service and the U.S. Forest Service by participating in reforestation and streambank stabilization projects, debris-flow monitoring and warning projects, as well as their ongoing role in snow-pack and water-supply forecasting. The National Weather Service continually provides short and long-term forecasts for precipitation and temperature. These figures are factored into projections for potential flooding, debris-flow, and fluctuating lake-level hazards. Expansion of the National Weather Service ALERT warning system has increased since 1982. The Department of Interior, through the Bureau of Reclamation, is involved in water control through the regulation of Federal dams for agricultural and flood control purposes. The Bureau of Reclamation is also currently repairing the Willard Reservoir embankments that were damaged by wave action on the Great Salt Lake. The U.S. Geological Survey is committing $3.2 million this fiscal year to seismic investigation as part of the National Earthquake Hazards Reduction Program. U.S. Fish and Wildlife is protecting what little of the Federal Bear River Migratory Bird Refuge that still remains above water. The Environmental Protection Agency, the Department of Housing and Urban Development, and the President's Jobs Bill have all contributed to hazard mitigation efforts in Utah. Other involvement of the Federal government includes NOAA and the Federal Highway Administration contributing the majority of funding for this workshop and a flood-warning workshop jointly sponsored by FEMA, NWS, and USGS last year. The speciality conference, "The Delineation of Landslide, Flash Flood, and Debris-flow Hazards in Utah" of last year was also sponsored, in part, by six Federal agencies.

* * *

Within FEMA itself, strong mitigation programs are beginning to be developed around the country. Among the available tools we have to help create opportunities for mitigation activities, is the Federal Interagency Hazard Mitigation Team. The Team is comprised of representatives from twelve Federal agencies with major authorities and programs that can be packaged for disaster preparedness, response, and recovery. **The Team has the responsibility of developing non-structural recommendations to Federal, State, and local governments, that will reduce the economic impact of future disasters on the Federal government.** As this process has evolved, the Team's efforts, more and more, are aimed at helping State and local levels of government help themselves. Although developed to produce a guidance report for Federal agencies immediately following a Presidential Declaration of a Major Disaster for floods, the Team was requested by Governor Matheson to survey the Utah Lake, Jordan River, Great Salt Lake system in February of 1983. These efforts were underway when the Thistle landslide diverted everyone's attention.

As a condition of future Federal disaster assistance, each State is now required to identify all of their hazards, assess their capability to respond to those events, and develop a plan to address any shortfalls that surfaced through this process. This long-term planning approach has become a valuable tool for FEMA nationwide in reducing the nation's vulnerability to hazards, and the financial burden they place on all levels of government. Failure to complete and aggressively address the recommendations of these plans can, and has, resulted in the delay or denial of Federal disaster assistance. The Utah State Hazard Mitigation Plan prepared in response to this requirement, is scheduled for completion in May.

The National Flood Insurance Program is the largest and best known program FEMA has to mitigate flood losses. While the program can be said to merely distribute the losses, the financial burden is subsidized primarily by those suffering the losses. Additionally, and equally as important, the program cleverly and intentionally makes this insurance available only after communities have agreed to adopt a minimum ordinance that severely restricts future development within identified flood hazard areas. Failure to properly enforce these ordinances can, and has, led to suspension from the Program, and thus denied the availability of insurance along with the permanent forms of disaster assistance involving structures or buildings.

Within the Public Assistance program of FEMA, that program which provides funding to State agencies and local political subdivisions for the repair or replacement of public facilities, there are provisions to complete this work with modifications intended to reduce the susceptability and impact from future disaster events.

All of these Federal activities also must comply with Executive Orders 11988 and 11990, which require all alternatives to be considered before a Federal agency can take any action that impacts a floodplain or wetland. If it is determined that an activity is one that might lead to supporting development in that area, for example the repair of streets and sewers, or the replacement of a bridge to a barrier island, the practicable alternatives must be considered outside of these areas before work can begin.

* * *

As outlined, there are currently many Federal activities underway, or of an ongoing nature, that contribute to our goals of mitigating the effects of future flooding, including the rise of the Great Salt Lake. In addition, and as a response to everyone's growing concerns, FEMA has stepped up their coordination role under our preliminary damage assessment responsibilities. We currently have people in Utah every other week monitoring the activities of all the Federal and State agencies responding to the Great Salt Lake problem. This information is developed into situation reports that are sent to the FEMA National Office in Washington, D.C., each Interagency Team member, each Senator and Congressman's office and the appropriate agencies within the State of Utah. Yesterday, the first meeting of the 1985 Utah Flood Task Force, comprised of Federal, State, and volunteer agencies, met at Utah's Division of Comprehensive Emergency Management. This Task Force will share information at the State level and provide for coordination between the Federal and State governments.

* * *

Most important at this point, though, is to take a step back and try to visualize all of these efforts together with those of the State. Both the Federal and State governments have a vested interest in what this conference produces. That explains why many of you are here through that endowment. For planning purposes, it is essential to be able to assimilate the best available data on what the Great Salt Lake level might do in order to finalize and justify the many pending decisions.

It is these pending decisions that I would like to address to complete this presentation. I want to reiterate, loud and clear, that all parties involved must not by-pass non-structural alternatives. Each and every day we read or hear about the alternatives being considered; the diking, raising, and protective work along the Interstate and railroads, the pumping of water into the western desert. Dikes and pumps. There may be nothing wrong with this approach. There may be nothing wrong with a "band-aid" response program (fix-it-as-it-breaks). But there also may be nothing wrong with doing nothing at all, or doing something totally different.

The frequency of people asking, "what if?" has diminished. What if the Interstate is raised again, and it becomes inundated again? What if it is raised again and the Lake recedes? What if the State spends $50 million to build a pumping station that is obsolete before it is completed?

We are caught in a hurry-up-we-have-to-do-something mentality. The logic being, if we don't do something now, then two years from now we will still be where we are today. I remind you of my opening quote from Mr. Austin. On the other hand, had something been done two years ago, maybe this feeling of panic wouldn't exist today.

There are still too many questions that haven't been answered. Why, now, is relocation of the Interstate just being given serious attention? Why is State and Federal money being used to subsidize housing in identified hazard areas? Was there any consideration given to the possible threat of flooding when constructing the new International Center or the expansion of the Salt Lake International Airport?

My point is, let us not forget our lessons. This is not the first time the Southern Pacific Railroad has had to face the possible loss of rail-lines due to an inland saltwater fluctuating lake. Their tracks bordering the Salton Sea have remained submerged since 1907. **It is time again to institute and implement long-term management strategies for the Great Salt Lake. This event, though unprecedented in its frequency to recur, should not be the problem it is. The problem is not that the Great Salt Lake goes up and down, or that Utah has experienced consecutive abnormal patterns of precipitation, but that Man has allowed development to occur where maybe he shouldn't have. We accept a certain degree of risk with everything we do. If what we are experiencing now is the risk accepted in the past, then fine, accept it. But let's not make the same mistake twice. Two years from now, or whenever the Lake recedes, a long-term strategy should be in place, and enforced. We should define what risk is acceptable to interstates, airports, wildlife areas, recreation facilities, private housing, sewage treatment plants, mineral extraction industries, business and industrial parks, and then direct our future efforts toward operating within those parameters.**

SESSION SUMMARIES AND RECOMMENDATIONS

SESSION SUMMARIES

Paleoenvironmental Record of Long-Term Fluctuations

George I. Smith, Rapporteur

The first five papers presented at this Conference provided a long-term perspective of the past fluctuations of Great Salt Lake, Lake Bonneville (its Pleistocene precursor), and the nature of the diverse climatic settings responsible for those fluctuations. The presentations left little doubt in my mind that both the former and present levels of this lake are expressions of climatic balances that prevailed for long periods over large areas; it must be more than coincidence that the climatically-controlled maximum areas of Pleistocene Lakes Bonneville and Lahontan (an approximately-contemporaneous lake that occupied much of northwest Nevada) were both about 17 times the areas of climatically-controlled present-lake areas. Fluctuations involving level changes of a few meters over periods lasting a few decades, like those recorded for Great Salt Lake since about 1847 to the present (Arnow, 1984), are probably also products of meteorologic variations, but their causes are inadequately understood.

The first four of those five papers reconstruct the history of lake-fluctuations or develop various climatic scenarios that might have produced one or more of the observed lake levels. Data for these studies came from (1) stratigraphic and geomorphic evidence along the basin's edges, (2) stable-isotope evidence from mid-lake sediments recovered as cores, (3) climatic constraints imposed by the hydrologic budgets of adjacent contemporaneous glaciers, and (4) reconstructed vegetation zones in the adjacent mountains and nearby valleys. The fifth paper summarizes the shorter but more recent climatic record preserved by rings in trees dating from the early 1700s.

These "long" records are relevant to the current problems addressed by the Conference in three ways. First, they show that variations in lake levels are the norm--changes that had many magnitudes and a variety of durations. Not only did the "largest" variations caused by climatic "change" take place over periods measured in "geologic" time, smaller variations--that were still large by the criteria in use by this conference--occurred over periods measured in a few hundred to a very few thousand years. These smaller fluctuations may have resulted from variations of something like 25 percent in one or more climatic components, and they are evident in both Pleistocene and Holocene records. Even smaller and briefer variations are evident in the detailed stable-isotope and tree-ring records which provide continuous records and numerical data amenable to statistical manipulation.

The second aspect of these studies that is relevant to this Conference is that the longer records, and the variety of approaches used to decipher them, emphasize the number of components of climate that can influence lake levels and help in ranking the potential importance of each. While opinions are divided, one school (to which I belong and which was supported by several Conference presentations) attributes the lake's most extreme fluctuation to variations in runoff volume, and substantially less to variations in factors that influence evaporation--air or water temperature, wind velocity, cloud cover, intensity amount, and atmospheric humidity. It is not that changes in the latter conditions are unable to produce variations in evaporation, but that most tend to be self-limiting. For example, a trend toward increased evaporation from a lake caused by warmer air, more wind, lower relative humidity, or increased radiation produced by decreased cloud cover causes a concurrent cooling of the lake water, decreased vapor pressure, and partial or complete termination of that trend. An opposite trend produces warmer water and more efficient evaporation. Increases in inflow, on the other hand, are self-limiting only to the extent that the inherent climate allows (record runoff volumes in from rivers in the Sierra Nevada (California) exceed 200 percent of "average" values).

The third relevant aspect of these papers to this Conference is the evidence of continuing lake-level change in the past which raises the question of whether the record of the past century can be considered one describing a "steady state." Statistical studies of what the past record promises for the future implicitly assume a steady state throughout both the past and future periods, and that might well be an unsafe practice. Prehistoric periods that seem to represent stable regimes lasted for centuries or a few millenia, but intuition suggests that the transitions between regimes--periods of non-steady states, by definition--probably required only years or decades. The probability of the last (or next) century including or being one of these transitional periods is small but not minute.

Two recommendations for future studies that might help remove part of the uncertainties of the future levels of Great Salt Lake:

(1) A series of shallow core holes should be drilled in the lowest-elevation parts of Great Salt Lake Desert Basin, the region first flooded whenever the Great Salt Lake level exceeded 4218 feet. This would help determine when and how frequently perennial lakes occupied that area over the last 5,000 to 10,000 years. Sediments deposited in perennial lakes are normally easily differentiated from those deposited in ephemeral or seasonal bodies of water, and age control (from carbon-14 data or secular paleomagnetic variation studies) should permit a well constrained history of such events. Knowledge of whether flow across this divide has happened, say, not even once within the last 5,000 years or several times during the last 500 years, should add (or remove) confidence in statistical estimates of such an event taking place in the foreseeable future. Flooding of the Great Salt Lake Desert basin as a consequence of inflow from its own, less mountainous watershed might have occurred periodically, but any persistent increase in regional precipitation would probably have also produced a rise in the Great Salt Lake itself with overflow into the western basin eventually resulting from it.

(2) A systematic study of the stable isotope content of the Wasatch Mountain snowpack during future years, at the end of the normal accumulation season, might help identify anomalies in that year's storm regimes. The conference attendees appeared to agree that anomalous but not unprecedented meteorological conditions seem to have been responsible for the heavy runoff in 1983 and 1984 which produced the unprecedented rises of the lake. The abundance of deuterium (hydrogen-2) in the Sierra Nevada snow pack has revealed year-to-year differences in seasonal weather patterns (Friedman and Smith, 1970, 1972, unpub. data), and variations in storm-tracks, moisture sources, and condensation temperatures appear to have contributed to these

differences (Smith and others, 1979). Some concern was expressed at the Conference as to whether different meteorological regimes, whose normal boundaries are not far from the Great Salt Lake watershed, might be migrating into the watershed and causing anomalous snowfall. The isotopic signatures of precipitation resulting from those regimes should be different, and those signatures could be used to identify and possibly understand the meteorological causes of anomalous seasons. It is unfortunate that such studies were not made during the anomalous winters that preceded the lake rise, but that cannot be remedied. Sampling the runoff from the season's snowpack, rather than the snowpack itself, was found in the Sierra Nevada studies (Smith and Friedman, unpub. data) to be unrepresentative of the snowpack, apparently because of large admixture of ground and soil water retained from previous years.

The uncertainty in the meteorological or climatic cause of the 1983 and 1984 rises of the Great Salt Lake, and the uncertainty as to what perspective should be adopted for purposes of planning, mean that long range projections must be made conservatively. The lake seems very unlikely to rise above the 4218-spillway level in the foreseeable future as its potential evaporative surface would increase by about a third once the spillway is reached. Salt Lake City is fortunate in that most of the city lies above 4218 feet, although those with property below that level will not feel that they share the city's good fortune. Nevertheless, from a perspective that is geological, yet not for periods too long to have human significance, I see no lower level that can be projected as a "maximum" level with any confidence or a compelling rationale. Even if that level is not reached in the near future, Salt Lake City with a probable future measured in centuries, should view areas below that level as more suitable for short-term or expendable purposes. Like the flood plains along large rivers and the supratidal zone along coastal plains, which are both within the natural purview of those water bodies, the zone below 4218 feet is--in a geologist's view--an encroachable zone because it is part of the Holocene floor of the Great Salt Lake.

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Historical Record Of Medium And Short-Term Fluctuations

Roger G. Barry, Rapporteur

Assessment of changes in lake level and their causes necessitates data on climatic and hydrologic variables. Streamflow contributes about two-thirds of the average annual inflow (3,576 x $10^6 m^3$) to Great Salt Lake and precipitation on the lake surface $\sim$ 31% (Arnow). The outflow is entirely by evaporation. Most of the streamflow originates from snowmelt in the Wasatch Range and Uinta Mountains to the east. Lake level is also affected by the basin hypsometry, consumptive water use, and a railroad causeway built in 1959, dividing the lake in two parts and thereby restricting brine flow and modifying the evaporation rate.

Observational data on lake level and climatic elements exist for just over one hundred years. Gage measurements of lake level began in 1875, with continuous recording since 1939. In addition, estimated values based on oral reports extend the data back to 1847. Most of the late nineteenth climatic records refer only to precipitation, with little temperature data available prior to 1910 (Bingham). Data from high elevation sites, where snowfall is predominant, are provided primarily from Soil Conservation Service snow courses, monitored monthly. Total loss of these valuable data is currently threatened by proposed funding cutbacks. Evaporation is measured by pan. Most streamflow records are from 1931.

Analysis of the annual September to May rise of lake level (Diaz and Kay) shows that it is well specified (r^2 = 0.78) by September-May precipitation in the Mountain and North Central Divisions combined with summer precipitation in the North-Central Division and summer cloudiness at Salt Lake City. The annual fall (June to August) of level is best predicted (r^2 = 0.78) by June-August precipitation in the North-Central Division and September-May precipitation in the Mountain Division. Evapotranspiration and temperature parameters do not show up in these regressions, perhaps due to multi-collinearities. Autumn precipitation, which contributes significantly to both sets of regressions, is subject to high variability. No significant periodicities are apparent, in the precipitation record, although a 2.0-2.2 year spectral peak in summer precipitation, and another peak in the neighborhood of 20 years was suggested.

Bradley's analysis of long-term precipitation data for the basin of Great Salt Lake shows that annual amounts of 150-175 percent of the 1921-70 mean occurred about 1905-10 and in the 1980s. For the October-September water year, 1981, 1982 and 1983 were among the top five wettest. The last few years were exceptionally wet in autumn, but also wet in spring and summer. Long-term trends in seasonal totals since the 1890s indicate higher values at both ends of the series, lower in the middle (except for winter precipitation), although the total variance accounted for by these trends is small.

Statistical estimates of return period (Karl and Young) indicate that the recent wet spell is "unusual, but not unexpected" in light of the precipitation climate of the last 100 years. The total of 1027 mm recorded in the Great Salt Lake Drainage between October 1981 and December 1984 has a return period between 118 and 290 years, depending on the data record ($>$100 years) used for the calculations. Derived precipitation series for the upper Bear River and the eastern and southern shores of Great Salt Lake appear to be stationary. However, streamflow estimates (Bowles and James) for 1862-72 are significantly different from the mean for 173-1983 or 1938-83, although the variance was (statistically) unchanged. It appears that the amounts have not been exceptional in the last few years, but the widespread nature of heavy falls over the entire basin and persistence over several years may have been unusual.

Streamflow data (1931-78) over the western United States show large-scale spatial coherence, with four principal component patterns (Lins). The level of the Great Salt Lake is correlated most highly ($r = 0.74$) with a spatial pattern that loads most highly in the Northern Great Plains. However, regression analysis of lake level with available tree-ring data suggests temporally and spatially varying relationships between climate and lake level (Bingham). Individual tree-ring sites give correlations with lake level comparable with those obtained from precipitation gage data for certain time intervals, but the site which provides the best explanation varies over time. Further analysis of the synoptic activity over the period of instrumental record seems desirable taking account of the spatial heterogeneity of precipitation regions in the Great Basin.

Physical Factors And Models Session #3

Colin S. Ramage, Rapporteur

Carpenter showed that the Great Salt Lake may enhance winter-time snow immediately downwind, about 6-8 times a year. The largest effect amounts to about 5-10% of annual precipitation and since it results from water evaporated from the lake that returns to the lake as run-off the net effect on lake level is negligible. Models based on statistical relationships among precipitation streamflow, evaporation and lake level use multiple regression (Stauffer), correlations (Williams), and stochastic simulations (Adams and associates, and Bowles and James) to exhibit some predictive skill out to about a year and progressively better predictions for shorter lead times. They raise the possibility of a graduated series of precautionary responses and place a premium on fast reaction time. Better specifications of evaporation and surface temperature and their lake-wide distribution could help improve the predictions. Perhaps a moored oceanographic buoy and some air-borne radiometric surface temperature measurements would be helpful as would a better understanding of synoptic patterns associated with significant lake-level fluctuations. Routine NOAA forecasts of Great Lakes levels (Croley) are also based on statistics and on weather analogs developed from past associations. The forecasts have some skill out to three weeks and might be modified for application to the Great Salt Lake. Finally, Geisler demonstrated no useful relationship between the El Nino cycle and conditions over the Great Salt Lake Basin.

And Now What?

Genevieve Atwood and Paul A. Kay, Rapporteurs

The last hour and a half of the conference was spent discussing the types of studies and research that could contribute to the understanding of those aspects of lake levels that we know we do not yet fully understand. Many of the conference participants spoke up and expressed the need for specific as well as more general types of information. The discussion ranged freely and there probably was less need for consensus on some studies than others but on four main areas for further study the group appeared to have reached consensus: 1) an endorsement of the idea of a political initiative in order to fund instrumentation of the Great Salt Lake drainage basin for climatic and water data; 2) careful examination of the geologic and archeologic evidence that could be collected to better understand the last 100 to 5000 years lake levels using shoreline information, geochemical information from cores, and volume calculations; 3) the need for socio-economic studies to determine the consequences to society of different levels of Great Salt Lake; 4) better information on climate factors from tree-ring studies and continued development of models.

Although some of the participants wanted to discuss various alternatives (structural and legislative) that could be used to diminish the damages from lake level rises or to control the level of the lake, the two moderators discouraged discussion of these two topics considering that this particular group might not be the best set of experts to discuss those topics, and for fear that the conference participants would not accomplish the stated goal of the session. . .to determine the need for future studies.

"PLANNING LEVEL" FOR THE LAKE

Although the conference participants were discouraged from enumerating the solutions to the problem of the Great Salt Lake level changes, the participants were encouraged by Bill Riebsame and Genevieve Atwood to define the nature of the problem itself; in particular, to define the "planning" level of the Great Salt Lake. Five alternatives for the planning level of the Great Salt Lake had been discussed during the previous session. These levels were: a level below 4212, 4212, 4217-4218 (the threshold level), 4222, and a level above 4222. The definition of "planning level" remained relatively obscure, and only loosely defined as that lake level which the participants would feel comfortable in advising policymakers to consider as a hazard. Phrased another way, it is the level above which the participants would not expect the lake to rise during the foreseeable lifetime of Salt Lake City. The "planning level" does not imply that the lake will rise to that level, simply that the possibility that it could rise to that level is significant enough that decisionmakers should factor it into their planning process. The 4212 level is the lake's historic high and has traditionally been considered the highest level to which the lake would probably rise. The 4217-4218 threshold level is the level at which the basin's geometry allows the lake to rapidly spread across more area with only a minor rise in elevation. This is the level at which the Great Salt Lake would naturally extend beyond the thresholds into the west desert. The 4222 level was discussed because of the probability as estimated by the Utah State University, Utah Water Research

Laboratory that there is a 3 percent chance that the lake could reach this level by the year 2050 AD.

There was considerable discussion on this topic and a universal agreement that levels below 4212 were too low to be considered as planning levels for the Great Salt Lake. Three participants felt that the 4212 level was sufficiently high, however the rest of the group felt it was "too low." Virtually all the participants except for those three considered the threshold level the best alternative for a planning level because of the basin's morphology. Judith McKenzie stated that her data show that during the last 500 years this level has been reached perhaps as many as five times and certainly a couple times. Most participants believed 4217-4218 is a "rational number to work with." One participant believed it is "too low" because only the Provo level (4780) could be considered a "truly safe number." The participants believed that the 4222 elevation was "not well enough defined," not rationally controlled by topography, and that the lake, although it had probably reached this level in the last 2000 years, had not reached it with the relative frequency as the threshold level.

The participants expressed many uncertainties about the 4217-4218 level. Not only is the threshold poorly defined topographically, members of the conference would have felt much more comfortable with that number if were known how often the lake had reached this level in the recent past and its persistence at that level. Many of the specific research topics suggested for further research would directly contribute to the better understanding of the threshold and its consequences to society.

GREAT BASIN INSTRUMENTATION INITIATIVE

Several participants were surprised and concerned to learn that the Soil Conservation Service's Snow Course program may be cut back or even eliminated due to financial constraints. This was highly disconcerting news to all participants. Virtually all participants agreed that the snow course network should be expanded and not allowed to be cut. Several participants suggested ideas on who to contact to keep the network funded. It was suggested that any cuts should be reinstated by western states governments if at all possible. It was clear that all participants believed in the necessity for good data base collection. The Snow Course data is one type of this information. Other types of basic data that had been used during the conference and some of which has been discontinued in recent years are stream gaging data collected by the United States Geological Survey's Water Resources Division and salinity and lake data collected by the Utah Geological and Mineral Survey. Basic data collection is expensive and endangered by budget constraints. Ted Arnow described how most individuals had been surprised when they learned that virtually every gaging station along the Wasatch Front had been discontinued and that virtually no real time data was available during the flooding along much of the Wasatch Front in the 1980s. The conference attendees clearly favored their continuance. The group also suggested that evaporation be measured from more than one station and that better techniques be developed for estimating evaporation loss.

There was a high degree of enthusiasm for the Great Salt Lake drainage basin instrumentation initiative and it is hoped that a group will further define the types of information that could best be collected on a regular basis through proper instrumentation in this area.

STUDIES TO FURTHER DEFINE LAKE LEVELS FOR THE LAST 5000 YEARS

Although near consensus had been reached on the "planning level" of the Great Salt Lake, there was obvious concern about the certainty with which we have determined past lake levels. Five types of studies were defined which would greatly enhance the understanding of frequency and duration of lake levels in the recent past.

1. It was suggested that cores could be taken in the deeper areas of the West Desert basin and analyzed geochemically and by sedimentological means to ascertain the frequency with which the west desert has been occupied by water. Participants felt that geochemical examination of the core could distinguish times when the West Desert basin and Great Salt Lake basin were united as a single lake. Sediments in the boggy areas along the edge of the playa could be analyzed for changes in organic constituents due to flooding of the west desert.

2. It was also suggested that additional core be taken in the Great Salt Lake basin and analyzed carefully for its geochemistry. Some of this work has already been done and has been shown to be a successful way to note changes in water level as reflected by salinity and carbonate deposition.

3. Further delineation of shore lines, particularly the lower most shore lines, and the application of archaeological findings to these shore lines has given some indication of flooded conditions in the past and appear to be a productive way to further delineate levels of the lake in the last 5000 years. Unlike the geochemical analysis of cores, these studies provide information about the actual dates of floods.

4. One of the most obviously needed pieces of information is the actual threshold level and detailed information concerning the basin morphology. The basin's geometry changes considerably at intervals between the levels 4215 and 4225 and this will effect evaporation. This is an important piece of information when determining the potential for the lake to stabilize naturally or artificially by West Desert pumping as precipitation and evaporation reach equalibrium. Most participants at the conference were surprised at the difficulty in determining the volume of the lake and correlating the volume of water with the lake level. Understanding the topography of the region would greatly assist in these volume calculations but it is also necessary to further identify rebound effects. All participants agreed that a detailed geodetic survey is a very high priority.

5. It was also suggested that the levels of other lakes in the Great Basin area be examined to see whether a pattern exists in the rise and fall regionally of lakes. Some participants cautioned that these studies could lead to false correlations and it was urged that the physical factors controlling these levels be identified and compared as well as the history of the levels of these lakes be examined.

CLIMATOLOGICAL STUDIES

There seems to be less agreement on the efficacy of further studies of climate and better estimates of the future levels of the lake. Climatologists are not confident of their ability to forecast longterm climate fluctuations, however their ability to predict in the shortterm (one to three months or even six months) has improved considerably. Several participants felt that only 20 more years of additional data would reveal whether this most recent period of above normal precipitation is a climate "change," or climate "fluctuation."

Virtually all participants agreed that tree-ring research is one area for further study. No tree-ring studies have been done in the Lake Basin to correlate climate with the level of the Great Salt Lake. Tree-ring studies done in the western United States have been more reliable in documenting dry periods than wet periods. It was suggested that trees that are known to be good indicators of wetter periods and which are located in the Great Salt Lake drainage basin itself be selected for a tree-ring study.

MODELS

Various models to best correlate past levels of the lake with climate had been presented during the conference and some models that were not presented by participants during the conference also were discussed during the summary session. Many participants felt that the probabilistic models had been fine tuned considerably. Conference participants expressed less confidence in the deterministic models. The public and decisionmakers undoubtedly have considerable difficulty distinguishing between probabilistic and deterministic models.

Most participants felt that further information would have to be collected before the present models would be improved considerably. There was a general agreement that a longer time frame would help most models' ability to predict future levels of the lake. Some aspects of the climatological models that could be further defined by more data collection include: basin geometry, evaporation, the formation and climatic consequences of winter ice on the lake, the potential for lake turnover as the lake freshens, precipitation and runoff into the Great Salt Lake from the west, spatial variability of precipitation, and the definition of climate sub-regimes and their consequence to the rise and fall of the Great Salt Lake.

SOCIO-ECONOMIC STUDIES

Participants agreed that the shortterm consequences of the different levels of the lake could be better defined and that the shortterm consequences of the structural responses that have been suggested should be further defined. The potential for catastrophic failure due to a malfunction of a dike or the probability of liquefaction from even a moderate earthquake was discussed but not in detail.

It was suggested that certain economic thresholds might be defined for policymakers as well as lake thresholds in order to better define the consequences of lake levels to policymakers.

CONCLUSION

The conference ended with thanks for the convenors. This apparently is the first time that the climatologists and the geologists have gotten together to discuss the levels of the Great Salt Lake and participants felt that the meeting had been highly productive and informational.

RECOMMENDATIONS

The recommendations that emerged at the conference, especially in the final meeting summarized by the rapporteurs, fall into two distinct, but not mutually exclusive, categories. A Great Salt Lake Basin instrumentation initiative was called for to ensure collection of real-time data needed by hydrological and climatological modellers. An integrated approach to deciphering the paleoenvironmental record would lead to answers concerning the timing and frequency of high and low lake levels, necessary for assessment of probabilistic statements about the current high levels. The recommendations are reiterated specifically below.

A. Real-Time Data

1. It is critical that the mountain snow course records be maintained; some of the relevant sites are outside the boundaries of the state.
2. Mountain precipitation data should be collected and used to assess the contribution of different altitudinal zones to streamflow.
3. Evaporation losses and precipitation input need to be measured directly over the lake at a central moored platform, as well as at other sites in the basin.
4. The climatological and synoptic aspects of short-term events, such as wind set-up of lake level and the distribution and movement of ice floes, require study through appropriate data collection.
5. Synoptic controls on the principal climatic and hydrologic variables need to be better determined.
6. Stable isotope analyses might provide information about the source regions of moisture and storm tracks, and should be determined for late-season snowpack in the Wasatch Mountains.

B. Paleoenvironmental Data

7. A series of shallow cores should be obtained from the deepest parts of the West Desert basin; sediment analyses would reveal timing and frequency of high lake levels (4218 feet and above).
8. Additional sediment analyses of cores from Great Salt Lake are needed to further elucidate the behavior of the lake, especially with careful dating control.
9. Further study of Holocene shorelines is needed to complete the record of lake level fluctuations; archaeological evidence would be of great importance in this endeavor.
10. Careful geodetic surveys are needed to delineate accurately the basin's morphology and to specify precisely the threshold to the West Desert.
11. Study of other lakes in the Great Basin should be done to detect whether there exist regional patterns of rise and fall.
12. Tree-ring studies are needed within the Great Salt Lake basin to provide reconstructions of climate and lake fluctuations on an annual basis for an additional several centuries prior to the historic record.

APPENDIX 1. DATA FOR GREAT SALT LAKE

The following tables present:

1. elevation-area-volume data for the whole lake, for the north arm, and for the south arm, for elevations between 4170 and 4218 feet, in whole and half foot increments;

2. maximum and minimum levels, and seasonal rises and falls, for the lake (measured at the south end) for the years 1851 to 1984;

3. monthly elevations on the first day of the month, for the period of measured record, 1875-1985 for the south arm of the lake and 1966-1985 for the north arm of the lake.

All elevation data have been reported in U.S. Geological Survey Water Supply Papers, as noted on the next page. The first two tables were provided by Norm Stauffer, Utah Division of Water Resources, and the third table and historical abstract of gage records by Ted Arnow, U.S. Geological Survey.

Records of surface elevation of Great Salt Lake

1847 - August 1874	Graph only based on reported information by G. K. Gilbert in Report on the lands of the Arid Region of the United States, by J. W. Powell, Washington, GPO, 1879, p. 64.
September 1875-January 1890	Periodic gage readings by various observers at Black Rock, Farmington, Lakeshore, and Garfield gages in Lake Bonneville, by G. K. Gilbert, U.S. Geol. Survey Monograph 1, 1890, p. 233-238. Summarized in U.S. Geol. Survey Water Supply Paper 1314, 1960, p. 24, and referenced to msl.
February 1890-December 1899	Periodic gage readings at a Garfield gage in U.S. Geol. Survey Water Supply Paper 38, 1900, p. 347-348. Summarized in U.S. Geol. Survey Water Supply Paper 1314, 1960, p. 24, and referenced to msl.
January 1900-September 1902	Graph only in USGS WSP 157, 1906, p. 26; also in USGS WSP 330, 1914, p. 16; also in USGS WSP 333, 1913, p. 15; also in USGS WSP 395, 1916, p. 195; also in USGS WSP 517, 1924, p. 16; also in USGS WSP 880,* 1941, p. 126. The graph may have been based on precipitation records or measurements made by the Oregon Short Line RR that were never published. On a print of a chart furnished by OSLRR, the records for Jan.-June 1900 and July 1901-September 1902 are shown by a solid line and the record for July 1900-June 1901 is shown by a dashed line.
October 1902-June 1903	Periodic gage readings by Southern Pacific RR summarized in USGS WSP 1314, 1960, p. 24, and referenced to msl.
July 1903-September 1938	Periodic gage readings at Saltair gage summarized in USGS WSP 1314, 1960, p. 24-26, and referenced to msl. Gage heights Mar. 9-July 21, 1904, in WSP 133. Since Oct. 1, 1912, gage heights have been published in annual water supply papers and the corresponding elevations have been published since 1958.
October 1939-	Continuous recorder readings at Salt Lake County boat harbor. Semimonthly records published in annual water suppl papers or State reports.

The graphs in WSP 330, 333, and 395 are identical; and the graphs in WSP 517 and 880 are identical and similar to the one in "Lands of the Arid Region of the U.S.", but they differ from the first group. The greatest variation is in the high stage period from 1866-75. The graph presently made by the USGS is that of the second group.

GREAT SALT LAKE ELEVATION-AREA-VOLUME DATA

	GREAT SALT LAKE		NORTH ARM		SOUTH ARM	
ELEVATION (FEET)	AREA (ACRES)	VOLUME (ACRE-FEET)	AREA (ACRES)	VOLUME (ACRE-FEET)	AREA (ACRES)	VOLUME (ACRE-FEET)
4170.0	118000.0	250000.0	36344.0	77000.0	81656.0	173000.0
4171.0	150000.0	384000.0	55000.0	122672.0	95000.0	261328.0
4172.0	180000.0	549000.0	70000.0	185172.0	110000.0	363828.0
4173.0	213000.0	745000.0	86000.0	263172.0	127000.0	481828.0
4174.0	244000.0	974000.0	98000.0	355172.0	146000.0	618828.0
4175.0	275000.0	1233500.0	112475.0	460410.0	162525.0	773090.0
4176.0	304000.0	1523000.0	115000.0	574148.0	189000.0	948852.0
4177.0	331000.0	1840498.0	129000.0	696148.0	202000.0	1144350.0
4178.0	358000.0	2184998.0	138000.0	829648.0	220000.0	1355350.0
4179.0	383000.0	2555498.0	146000.0	971648.0	237000.0	1583850.0
4180.0	407000.0	2950500.0	152625.0	1120960.0	254375.0	1829540.0
4181.0	427000.0	3367480.0	160000.0	1277270.0	267000.0	2090210.0
4182.0	446000.0	3803990.0	166000.0	1440270.0	280000.0	2363720.0
4183.0	465000.0	4259490.0	172000.0	1609270.0	293000.0	2650220.0
4184.0	482000.0	4732990.0	179000.0	1784770.0	303000.0	2948220.0
4185.0	496124.0	5222050.0	187624.0	1968080.0	308500.0	3253970.0
4186.0	509034.0	5724620.0	193000.0	2158390.0	316034.0	3566230.0
4187.0	522000.0	6240140.0	198000.0	2353890.0	324000.0	3886250.0
4188.0	535056.0	6768670.0	203000.0	2554390.0	332056.0	4214280.0
4189.0	550000.0	7311200.0	207000.0	2759390.0	343000.0	4551810.0
4190.0	564196.0	7868300.0	212954.0	2969370.0	351242.0	4898930.0
4190.5	572094.0	8154350.0	214973.0	3076860.0	357121.0	5077490.0
4191.0	580000.0	8440400.0	217000.0	3184350.0	363000.0	5256050.0
4191.5	589400.0	8732750.0	220000.0	3293600.0	369400.0	5439150.0
4192.0	601861.0	9030560.0	227000.0	3405350.0	374861.0	5625210.0
4192.5	613500.0	9334400.0	233000.0	3520350.0	380500.0	5814050.0
4193.0	632676.0	9645950.0	246730.0	3640280.0	385946.0	6005670.0
4193.5	655000.0	9967860.0	252000.0	3764960.0	403000.0	6202900.0
4194.0	677888.0	10301090.0	261300.0	3893290.0	416588.0	6407800.0
4194.5	703000.0	10646310.0	270000.0	4026110.0	433000.0	6620200.0

GREAT SALT LAKE ELEVATION-AREA-VOLUME DATA

	GREAT SALT LAKE		NORTH ARM		SOUTH ARM	
ELEVATION (FEET)	AREA (ACRES)	VOLUME (ACRE-FEET)	AREA (ACRES)	VOLUME (ACRE-FEET)	AREA (ACRES)	VOLUME (ACRE-FEET)
4195.0	719964.0	11002040.0	276360.0	4162700.0	443604.0	6839340.0
4195.5	748900.0	11369270.0	287000.0	4303540.0	461900.0	7065730.0
4196.0	772964.0	11749730.0	298580.0	4449940.0	474384.0	7299790.0
4196.5	807000.0	12144720.0	310000.0	4602080.0	497000.0	7542640.0
4197.0	839809.0	12556430.0	326100.0	4761110.0	513709.0	7795320.0
4197.5	866000.0	12982870.0	335000.0	4926380.0	531000.0	8056490.0
4198.0	890047.0	13421890.0	343720.0	5096060.0	546327.0	8325830.0
4198.5	926000.0	13875900.0	360000.0	5271990.0	566000.0	8603910.0
4199.0	969949.0	14350140.0	382220.0	5457800.0	587729.0	8892340.0
4199.5	1015000.0	14846630.0	390000.0	5651100.0	625000.0	9195530.0
4200.0	1079259.0	15370180.0	405000.0	5849850.0	674259.0	9520330.0
4200.5	1113000.0	15918250.0	415000.0	6054850.0	698000.0	9863400.0
4201.0	1140000.0	16481450.0	425000.0	6264850.0	715000.0	10216600.0
4201.5	1161000.0	17056750.0	435000.0	6479850.0	726000.0	10576900.0
4202.0	1175000.0	17640700.0	440000.0	6698600.0	735000.0	10942100.0
4202.5	1188000.0	18231450.0	445000.0	6919850.0	743000.0	11311600.0
4203.0	1201000.0	18828700.0	450000.0	7143600.0	751000.0	11685100.0
4203.5	1212000.0	19431950.0	455000.0	7369850.0	757000.0	12062100.0
4204.0	1223000.0	20040700.0	460000.0	7598600.0	763000.0	12442100.0
4204.5	1235000.0	20655250.0	465000.0	7829850.0	770000.0	12825400.0
4205.0	1250468.0	21275600.0	470000.0	8063600.0	780468.0	13212000.0
4205.5	1290234.0	21908750.0	472272.0	8306550.0	817962.0	13602200.0
4206.0	1330000.0	22541900.0	474544.0	8549500.0	855456.0	13992400.0
4206.5	1352500.0	23175150.0	476816.0	8792450.0	875684.0	14382700.0
4207.0	1375000.0	23808300.0	479088.0	9035400.0	895912.0	14772900.0
4207.5	1392500.0	24441450.0	481360.0	9278350.0	911140.0	15163100.0
4208.0	1410000.0	25074700.0	483632.0	9521300.0	926368.0	15553400.0
4208.5	1430000.0	25707850.0	485904.0	9764250.0	944096.0	15943600.0
4209.0	1450000.0	26341000.0	488176.0	10007200.0	961824.0	16333800.0
4209.5	1470000.0	26974200.0	490448.0	10250100.0	979552.0	16724100.0
4210.0	1490000.0	27607300.0	492720.0	10493000.0	997280.0	17114300.0
4212.0	1572000.0	30669000.0	534000.0	11520000.0	1038000.0	19149000.0
4216.0	2228000.0	38671000.0	1121000.0	15233000.0	1107000.0	23438000.0
4218.0	2519000.0	43417000.0	1372000.0	17726000.0	1147000.0	25691000.0

GREAT SALT LAKE

YEAR	HIGH (FEET)	LOW (FEET)	RISE (FEET)	FALL (FEET)
1851	4202.10	4200.80	1.10	1.30
1852	4203.00	4201.90	2.20	1.10
1853	4204.00	4203.00	2.10	1.00
1854	4204.20	4203.60	1.20	0.60
1855	4204.60	4203.00	1.00	1.60
1856	4204.20	4203.10	1.20	1.10
1857	4204.00	4202.00	0.90	2.00
1858	4203.00	4201.00	1.00	2.00
1859	4201.80	4200.20	0.80	1.60
1860	4201.30	4199.60	1.10	1.70
1861	4201.00	4199.60	1.40	1.40
1862	4203.00	4201.80	3.40	1.20
1863	4204.00	4202.80	2.20	1.20
1864	4204.50	4203.60	1.70	0.90
1865	4205.40	4204.90	1.80	0.50
1866	4207.00	4206.40	2.10	0.60
1867	4208.40	4207.70	2.00	0.70
1868	4210.00	4209.30	2.30	0.70
1869	4211.00	4209.90	1.70	1.10
1870	4210.60	4209.20	0.70	1.40
1871	4210.50	4209.60	1.30	0.90
1872	4211.40	4210.50	1.80	0.90
1873	4211.60	4210.60	1.10	1.00
1874	4211.30	4209.70	0.70	1.60
1875	4211.00	4208.60	1.30	2.40
1876	4210.90	4209.90	2.30	1.00
1877	4210.40	4208.90	0.50	1.50
1878	4209.40	4207.80	0.50	1.60
1879	4208.10	4205.60	0.30	2.50
1880	4206.50	4204.80	0.90	1.70
1881	4206.50	4205.10	1.70	1.40
1882	4206.00	4204.50	0.90	1.50
1883	4205.00	4203.50	0.50	1.50
1884	4205.90	4205.30	2.40	0.60
1885	4207.30	4206.20	2.00	1.10
1886	4207.70	4206.50	1.50	1.20
1887	4207.20	4205.60	0.70	1.60
1888	4206.10	4203.90	0.50	2.20
1889	4204.80	4202.10	0.90	2.70
1890	4204.10	4202.70	2.00	1.40
1891	4203.50	4202.30	0.80	1.20
1892	4202.90	4201.60	0.60	1.30
1893	4203.00	4201.00	1.40	2.00
1894	4203.00	4201.60	2.00	1.40
1895	4202.20	4200.70	0.60	1.50
1896	4201.80	4200.40	1.10	1.40
1897	4202.30	4200.50	1.90	1.80
1898	4201.90	4199.80	1.40	2.10
1899	4201.20	4199.90	1.40	1.30
1900	4201.00	4198.90	1.10	2.10

GREAT SALT LAKE

YEAR	HIGH (FEET)	LOW (FEET)	RISE (FEET)	FALL (FEET)
1901	4200.00	4198.40	1.10	1.60
1902	4199.00	4196.50	0.60	2.50
1903	4197.70	4196.10	1.20	1.60
1904	4198.30	4196.90	2.20	1.40
1905	4197.60	4195.80	0.70	1.80
1906	4198.30	4197.10	2.50	1.20
1907	4200.50	4199.80	3.40	0.70
1908	4201.00	4199.90	1.20	1.10
1909	4202.60	4202.00	2.70	0.60
1910	4203.90	4201.90	1.90	2.00
1911	4203.20	4201.60	1.30	1.60
1912	4202.70	4201.80	1.10	0.90
1913	4202.90	4201.70	1.10	1.20
1914	4203.60	4202.40	1.90	1.20
1915	4203.20	4201.40	0.80	1.80
1916	4202.90	4201.30	1.50	1.60
1917	4203.40	4202.30	2.10	1.10
1918	4203.50	4201.90	1.20	1.60
1919	4202.70	4200.70	0.80	2.00
1920	4202.00	4200.70	1.30	1.30
1921	4203.30	4201.90	2.60	1.40
1922	4204.40	4202.90	2.50	1.50
1923	4204.90	4203.70	2.00	1.20
1924	4205.10	4203.00	1.40	2.10
1925	4204.30	4203.30	1.30	1.00
1926	4204.30	4202.40	1.00	1.90
1927	4203.50	4201.80	1.10	1.70
1928	4202.60	4200.70	0.80	1.90
1929	4202.00	4200.25	1.30	1.75
1930	4201.20	4199.70	0.95	1.50
1931	4200.45	4197.90	0.75	2.55
1932	4199.40	4197.75	1.50	1.65
1933	4198.85	4196.75	1.10	2.10
1934	4197.20	4194.80	0.45	2.40
1935	4196.05	4193.85	1.25	2.20
1936	4195.80	4194.35	1.95	1.45
1937	4196.45	4194.80	2.10	1.65
1938	4196.55	4194.90	1.75	1.65
1939	4196.55	4194.60	1.65	1.95
1940	4195.75	4193.70	1.15	2.05
1941	4195.65	4194.05	1.95	1.60
1942	4196.55	4194.65	2.50	1.90
1943	4196.20	4194.60	1.55	1.60
1944	4196.45	4194.85	1.85	1.60
1945	4196.40	4195.20	1.55	1.20
1946	4197.15	4195.40	1.95	1.75
1947	4197.20	4196.25	1.80	0.95
1948	4197.75	4196.15	1.50	1.60
1949	4198.25	4196.70	2.10	1.55
1950	4198.80	4197.55	2.10	1.25

GREAT SALT LAKE

YEAR	HIGH (FEET)	LOW (FEET)	RISE (FEET)	FALL (FEET)
1951	4199.90	4198.35	2.35	1.55
1952	4200.95	4199.35	2.60	1.60
1953	4200.55	4198.75	1.20	1.80
1954	4199.35	4197.15	0.60	2.20
1955	4198.05	4196.40	0.90	1.65
1956	4197.85	4195.90	1.45	1.95
1957	4197.45	4195.85	1.55	1.60
1958	4197.40	4195.25	1.55	2.15
1959	4196.05	4194.30	0.80	1.75
1960	4195.30	4193.20	1.00	2.10
1961	4193.80	4191.60	0.60	2.20
1962	4193.85	4191.90	2.25	1.95
1963	4192.95	4191.35	1.05	1.60
1964	4194.15	4192.40	2.80	1.75
1965	4194.25	4193.65	1.85	0.60
1966	4195.60	4193.20	1.95	2.40
1967	4195.15	4193.80	1.95	1.35
1968	4195.60	4194.20	1.80	1.40
1969	4197.10	4195.10	2.90	2.00
1970	4196.40	4194.85	1.30	1.55
1971	4198.00	4196.85	3.15	1.15
1972	4199.60	4197.90	2.75	1.70
1973	4200.45	4199.15	2.55	1.30
1974	4201.30	4199.15	2.15	2.15
1975	4201.55	4199.90	2.40	1.65
1976	4202.25	4200.30	2.35	1.95
1977	4200.75	4198.65	0.45	2.10
1978	4200.25	4198.40	1.60	1.85
1979	4199.90	4197.50	1.50	2.40
1980	4200.55	4199.00	3.05	1.55
1981	4200.15	4198.15	1.15	2.00
1982	4200.70	4199.65	2.55	1.05
1983	4204.70	4204.30	5.05	0.40
1984	4209.25	4207.85	4.95	1.40
MEAN	4201.77	4200.22	1.60	1.55
STD DEV	4.44	4.61	0.81	0.48
STD ERR	0.38	0.40	0.07	0.04

SOUT H ARM   GREAT SALT LAKE LEVELS (DATUM = 4000 FT ABOVE SEA LEVEL

YEAR	DAY	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
1875	1	M	M	M	M	M	M	M	M	208.90	208.70	208.60	208.80
1876	1	209.00	209.10	209.20	209.40	209.70	210.30	210.90	210.60	210.00	209.90	M	M
1877	1	M	M	M	M	M	M	210.40	M	M	M	210.20	208.90
1878	1	M	209.00	M	209.10	209.30	M	209.40	209.20	M	M	207.90	207.80
1879	1	M	M	M	M	208.10	M	M	M	M	M	M	205.60
1880	1	205.80	205.70	205.80	205.90	206.10	204.40	206.50	206.20	205.80	205.10	205.00	204.80
1881	1	204.90	205.10	205.60	205.70	206.10	206.50	206.30	206.00	205.50	205.20	205.10	205.10
1882	1	205.20	205.30	205.40	205.70	205.90	206.00	205.80	205.50	204.90	204.60	204.50	204.50
1883	1	204.50	204.50	204.60	204.70	M	M	M	M	205.00	204.10	203.70	203.50
1884	1	203.50	203.50	203.70	204.10	204.70	205.50	205.90	205.70	205.50	205.50	205.40	205.30
1885	1	205.60	205.80	206.00	206.30	206.60	207.00	207.30	207.10	206.70	206.50	206.40	206.20
1886	1	206.50	206.80	207.10	207.30	207.50	207.70	M	207.30	M	206.70	206.50	M
1887	1	206.70	206.60	206.80	206.90	206.90	207.20	207.00	206.60	206.20	205.80	205.60	M
1888	1	205.60	205.70	205.80	206.10	205.90	206.00	205.60	205.20	204.80	204.40	203.90	204.10
1889	1	204.10	204.20	204.50	204.80	204.40	204.20	203.70	203.10	202.60	202.10	M	202.20
1890	1	202.30	202.70	203.00	203.30	203.70	M	204.10	203.70	203.20	M	202.70	202.70
1891	1	202.80	202.80	203.10	203.20	203.40	203.50	M	M	202.80	202.40	202.40	202.30
1892	1	202.30	M	M	M	M	M	M	M	M	M	M	M
1893	1	201.70	201.90	202.00	202.40	202.80	203.00	202.80	202.00	201.80	201.00	M	M
1894	1	201.50	201.70	201.60	202.10	202.50	203.00	M	M	M	201.90	201.60	M
1895	1	M	M	M	M	M	M	M	M	M	M	M	M
1896	1	200.80	201.00	201.10	201.30	201.30	201.80	201.80	201.30	200.90	200.70	200.40	200.50
1897	1	200.80	201.00	201.30	201.60	201.90	202.30	202.00	201.50	201.20	200.70	200.50	200.60
1898	1	200.80	200.90	201.20	201.40	201.70	201.90	201.80	201.30	200.70	200.30	200.00	199.80
1899	1	199.90	200.20	200.50	200.80	200.90	201.20	201.20	200.80	200.50	200.20	200.20	199.90
1900	1	M	M	M	M	M	M	M	M	M	M	M	M
1901	1	M	M	M	M	M	M	M	M	M	M	M	M
1902	1	M	M	M	M	M	M	M	M	M	196.90	196.50	196.60
1903	1	196.70	196.90	197.00	197.30	197.60	197.70	197.70	197.10	196.70	196.20	196.10	196.10
1904	1	196.40	196.40	196.80	197.60	197.70	198.30	198.30	197.90	197.50	197.20	196.90	196.90
1905	1	196.10	197.00	197.20	197.50	197.60	197.60	197.30	196.80	196.30	196.00	195.80	195.90
1906	1	195.90	196.10	196.40	197.00	197.40	197.90	198.30	197.80	197.50	197.30	197.10	197.10
1907	1	197.40	197.70	198.40	198.90	199.50	199.90	200.50	200.40	200.10	199.80	199.90	199.90
1908	1	200.10	200.30	200.60	200.70	200.70	200.80	201.00	200.70	200.10	199.90	200.00	200.10
1909	1	200.30	200.80	201.10	201.50	201.70	202.40	202.60	202.40	202.10	202.00	202.00	202.30
1910	1	202.60	202.90	203.10	203.60	203.90	203.90	203.40	203.00	202.50	202.10	202.00	201.90
1911	1	202.40	202.30	202.70	203.00	203.20	203.10	203.00	202.70	202.10	201.60	201.70	201.60
1912	1	201.60	201.80	202.00	202.20	202.50	202.60	202.70	202.40	202.10	201.80	201.90	202.10
1913	1	202.10	202.10	202.30	202.60	202.80	202.90	202.80	202.50	202.00	201.90	201.70	201.80
1914	1	201.90	202.30	202.60	202.80	203.10	203.30	203.60	203.30	202.80	202.40	202.50	202.40
1915	1	202.50	202.60	202.90	203.10	203.20	203.20	202.90	202.40	201.90	201.60	201.40	201.40
1916	1	201.40	201.60	202.10	202.50	202.90	202.80	202.60	202.20	201.60	201.30	201.40	201.40
1917	1	201.50	201.60	201.90	202.30	202.80	203.10	203.40	203.20	202.70	202.50	202.30	202.50
1918	1	202.60	202.90	203.10	203.50	203.50	203.40	203.10	202.90	202.40	202.10	201.90	202.00

SOUT H ARM GREAT SALT LAKE LEVELS (DATUM = 4000 FT ABOVE SEA LEVEL

YEAR	DAY	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
1919	1	202.20	202.30	202.50	202.70	202.70	202.60	202.00	201.50	201.10	200.80	200.70	200.70
1920	1	200.80	200.90	201.20	201.40	201.70	202.00	201.90	201.50	201.10	200.70	200.90	201.10
1921	1	201.30	201.50	201.80	202.30	202.70	203.20	203.30	202.80	202.20	201.90	201.90	202.00
1922	1	202.30	202.50	202.90	203.30	203.70	204.30	204.40	203.90	203.60	203.10	202.90	203.10
1923	1	203.30	203.70	203.80	204.10	204.40	204.90	204.80	204.40	204.10	203.70	204.00	204.10
1924	1	204.20	204.40	204.70	204.90	205.10	204.70	204.50	204.00	203.40	203.10	203.00	203.10
1925	1	203.30	203.50	203.80	204.10	204.30	204.30	204.10	203.90	203.50	203.30	203.30	203.40
1926	1	203.50	203.60	203.90	204.10	204.30	204.20	203.80	203.40	202.90	202.50	202.40	202.40
1927	1	202.50	202.60	202.90	203.30	203.40	203.50	203.20	202.80	202.10	201.90	201.80	202.00
1928	1	202.00	202.10	202.30	202.60	202.60	202.50	202.20	201.70	201.80	200.70	200.70	200.70
1929	1	200.90	201.00	201.10	201.60	201.90	202.00	201.70	201.20	200.80	200.65	200.25	200.45
1930	1	200.55	200.70	200.90	201.10	201.15	201.20	200.85	200.30	200.05	199.70	199.85	200.00
1931	1	200.10	200.15	200.45	200.40	200.35	200.20	199.60	199.20	198.65	198.05	198.00	197.90
1932	1	198.10	198.20	198.55	198.85	199.10	199.40	199.25	198.85	198.35	198.00	197.80	197.75
1933	1	197.70	197.95	198.10	198.40	198.55	198.85	198.60	198.10	197.45	197.05	196.80	196.75
1934	1	196.80	197.05	197.20	197.20	197.10	196.60	196.25	195.85	195.35	195.00	194.80	194.95
1935	1	195.15	195.30	195.55	195.70	195.75	196.05	195.70	195.15	194.60	194.20	193.85	193.90
1936	1	194.00	194.25	194.80	195.05	195.50	195.80	195.75	195.40	194.80	194.35	194.35	194.55
1937	1	194.65	194.85	195.30	196.10	196.40	196.45	196.20	195.90	195.30	194.95	194.80	194.80
1938	1	195.10	195.25	195.55	195.95	196.20	196.55	196.20	195.95	195.45	194.90	195.05	195.20
1939	1	195.50	195.70	195.90	196.30	196.55	196.25	195.95	195.45	194.95	194.75	194.65	194.60
1940	1	194.65	195.10	195.50	195.70	195.75	195.40	194.95	194.45	193.90	193.80	193.70	193.90
1941	1	194.30	194.50	195.05	195.25	195.65	195.55	195.30	194.85	194.40	194.05	194.25	194.40
1942	1	194.70	195.05	195.50	196.00	196.30	196.55	196.30	195.80	195.25	194.85	194.65	194.80
1943	1	195.10	195.30	195.65	195.85	196.20	196.20	196.05	195.60	195.05	194.75	194.60	194.75
1944	1	195.05	195.15	195.50	195.90	196.20	196.30	196.45	196.00	195.35	194.95	194.85	194.90
1945	1	195.05	195.20	195.65	195.80	195.90	196.00	196.40	195.90	195.60	195.20	195.25	195.50
1946	1	195.75	196.05	196.20	196.70	197.10	197.15	196.85	196.35	195.85	195.40	195.55	195.85
1947	1	196.25	196.35	196.90	197.00	197.20	197.15	197.15	196.80	196.35	196.25	196.25	196.50
1948	1	196.70	196.85	197.05	197.30	197.60	197.75	197.75	197.15	196.55	196.25	196.15	196.25
1949	1	196.50	196.75	197.05	197.80	198.00	198.25	198.00	197.50	197.05	196.70	196.80	197.00
1950	1	197.15	197.40	197.90	198.15	198.35	198.80	198.65	198.30	197.90	197.55	197.60	197.90
1951	1	198.30	198.50	198.95	199.20	199.45	199.90	199.55	199.15	198.85	198.45	198.35	198.55
1952	1	198.80	199.20	199.50	200.05	200.60	200.95	200.75	200.40	199.90	199.55	199.40	199.35
1953	1	199.60	200.05	200.20	200.30	200.50	200.55	200.50	200.00	199.35	198.95	198.75	198.75
1954	1	198.80	199.05	199.15	199.30	199.35	199.05	198.85	198.35	197.70	197.35	197.15	197.25
1955	1	197.30	197.45	197.60	197.85	198.05	197.95	197.80	197.30	196.95	196.55	196.40	196.45
1956	1	196.85	197.30	197.50	197.60	197.75	197.85	197.50	197.05	196.35	196.00	195.90	195.90
1957	1	196.15	196.30	196.55	196.75	197.00	197.45	197.40	196.95	196.30	196.00	195.85	195.95
1958	1	196.15	196.35	196.80	196.95	197.40	197.40	196.95	196.35	195.90	195.50	195.25	195.30
1959	1	195.50	195.75	195.95	196.00	196.05	195.95	195.70	195.30	194.80	194.50	194.40	194.30
1960	1	194.50	194.65	194.85	195.30	195.30	195.15	194.75	194.25	193.65	193.35	193.20	193.20
	15	M	M	M	M	M	M	M	M	M	193.30	193.20	193.30
1961	1	193.35	193.40	193.60	193.80	193.65	193.40	193.05	192.45	191.90	191.70	191.60	191.75

GREAT SALT LAKE LEVELS (DATUM = 4000 FT ABOVE SEA LEVEL

SOUT H ARM

YEAR	DAY	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
1961	15	193.35	193.50	193.70	193.75	193.55	193.25	192.80	192.25	191.70	191.70	191.60	191.85
1962	1	192.10	192.15	192.95	193.30	193.60	193.80	193.70	193.10	192.55	192.15	191.90	191.90
	15	192.10	192.25	193.10	193.40	193.65	193.85	193.35	192.85	192.30	191.90	191.90	191.90
1963	1	191.95	191.95	192.35	192.40	192.75	192.80	192.75	192.15	191.65	191.50	191.35	191.70
	15	191.95	192.25	192.35	192.55	192.85	192.95	192.55	191.90	191.55	191.35	191.45	191.75
1964	1	191.85	191.90	192.05	192.45	193.00	193.45	194.15	193.60	193.00	192.60	192.40	192.40
	15	191.85	192.05	192.10	192.75	193.35	193.85	194.00	193.35	192.80	192.50	192.40	192.50
1965	1	192.90	193.40	193.65	193.70	194.05	194.25	194.20	193.85	193.65	193.75	193.80	194.10
	15	193.15	193.55	193.70	193.80	194.25	194.20	194.05	193.75	193.75	193.80	193.85	194.40
1966	1	194.40	194.75	195.10	195.55	195.60	195.40	194.90	194.35	193.75	193.40	193.20	193.30
	15	194.70	194.95	195.45	195.60	195.55	195.15	194.55	194.05	193.45	193.30	193.30	193.40
1967	1	193.50	193.70	193.95	194.05	194.30	194.45	195.15	194.65	194.15	193.85	193.80	193.90
	15	193.50	193.85	193.95	194.10	194.40	194.85	194.95	194.40	193.95	193.80	193.85	193.90
1968	1	194.00	194.25	194.80	195.30	195.45	195.40	195.45	194.95	194.65	194.30	194.40	194.55
	15	194.15	194.35	195.15	195.40	195.40	195.60	195.20	194.70	194.55	194.20	194.45	194.70
1969	1	194.90	195.50	196.05	196.70	197.05	196.90	196.65	196.30	195.70	195.25	195.25	195.45
	15	195.10	195.70	196.30	197.00	197.10	196.70	196.55	196.00	195.45	195.10	195.35	195.55
1970	1	195.70	196.10	196.30	196.40	196.35	196.40	196.25	195.70	195.25	194.90	194.90	195.25
	15	195.70	196.25	196.35	196.30	196.35	196.30	196.05	195.50	195.10	194.85	195.15	195.50
1971	1	195.65	196.25	196.70	196.95	197.55	197.90	198.00	197.55	197.05	196.85	197.15	197.60
	15	195.80	196.50	196.80	197.10	197.75	198.00	197.80	197.25	197.05	197.00	197.45	197.65
1972	1	197.85	198.25	198.70	199.00	199.50	199.50	199.15	198.60	198.10	197.90	198.10	198.40
	15	198.00	198.40	198.90	199.25	199.60	199.40	199.00	198.40	197.95	197.95	198.25	198.45
1973	1	198.65	198.95	199.35	200.15	200.35	200.45	200.15	199.70	199.15	199.20	199.20	199.35
	15	198.75	199.20	199.80	200.20	200.45	200.35	199.85	199.45	199.15	199.20	199.25	199.45
1974	1	199.60	199.85	200.20	200.70	201.15	201.30	201.00	200.20	199.65	199.25	199.30	199.35
	15	199.65	200.05	200.55	201.05	201.25	201.25	200.45	199.85	199.35	199.15	199.35	199.45
1975	1	199.55	199.65	199.95	200.45	200.90	201.35	201.45	201.05	200.40	200.00	200.10	200.30
	15	199.55	199.80	200.20	200.60	201.15	201.55	201.40	200.75	200.20	199.90	200.20	200.45
1976	1	200.60	200.80	201.20	201.65	202.10	202.25	201.85	201.25	200.70	200.40	200.30	200.30
	15	200.70	201.10	201.50	201.85	202.20	202.00	201.55	200.90	200.50	200.40	200.35	200.40
1977	1	200.45	200.50	200.65	200.75	200.65	200.70	200.25	199.65	199.25	198.90	198.70	198.65
	15	200.50	200.60	200.75	200.75	200.50	200.60	199.95	199.40	199.00	198.80	198.65	198.70
1978	1	198.80	199.10	199.40	199.80	200.20	200.25	199.85	199.25	198.70	198.55	198.40	198.50
	15	198.85	199.30	199.65	199.95	200.20	200.20	199.60	198.90	198.45	198.45	198.40	198.55
1979	1	198.65	198.75	199.10	199.75	199.85	199.75	199.25	198.65	198.15	197.65	197.60	197.60
	15	198.70	198.90	199.40	199.85	199.90	199.50	198.90	198.35	197.85	197.50	197.55	197.65
1980	1	197.70	198.25	198.85	199.20	199.40	200.10	200.35	199.90	199.30	199.10	199.00	199.25
	15	197.85	198.35	199.10	199.30	199.85	200.55	200.15	199.40	199.20	199.00	199.10	199.35
1981	1	199.50	199.80	199.85	200.00	200.00	200.05	199.90	199.20	198.60	198.15	198.35	198.40
	15	199.65	199.85	199.95	200.05	199.90	200.15	199.55	198.85	198.40	198.35	198.40	198.45
1982	1	198.60	198.80	199.20	199.80	200.25	200.70	200.45	200.25	199.75	200.05	200.50	200.90
	15	198.70	198.85	199.45	200.10	200.55	200.70	200.35	199.90	199.65	200.30	200.70	201.25

GREAT SALT LAKE LEVELS (DATUM = 4000 FT ABOVE SEA LEVEL

SOUT H ARM

YEAR	DAY	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
1983	1	201.45	201.90	202.25	203.00	203.50	204.05	204.70	204.45	204.40	204.35	204.70	205.05
	15	201.60	202.15	202.45	203.25	203.65	204.50	204.65	204.30	204.40	204.55	204.75	205.40
1984	1	205.90	206.30	206.70	207.35	208.10	208.80	209.25	209.05	208.15	207.85	208.05	208.30
	15	206.15	206.50	206.90	207.75	208.35	209.15	209.10	208.55	208.00	207.90	208.20	208.45
1985	1	208.65	208.90	209.15									
	15	208.80	209.00	209.35									

NORT H ARM GREAT SALT LAKE LEVELS (DATUM = 4000 FT ABOVE SEA LEVEL

YEAR	DAY	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
1966	1	M	M	M	M	194.70	194.70	193.10	193.65	193.10	192.85	192.65	192.75
	15	M	M	M	194.70	194.70	194.40	193.35	193.40	193.00	192.65	192.75	193.10
1967	1	192.85	M	193.30	193.45	193.55	193.75	194.15	M	193.50	193.20	193.10	193.20
	15	192.90	M	193.40	193.55	193.70	194.00	194.10	193.70	193.35	193.15	193.20	193.20
1968	1	193.30	193.55	193.90	194.40	194.45	194.45	194.45	194.05	193.85	193.55	193.55	193.65
	15	193.40	193.70	194.15	194.45	194.50	194.60	194.35	193.95	193.70	193.45	193.60	193.75
1969	1	194.00	194.55	195.10	195.45	195.75	195.70	195.50	195.10	194.60	194.30	194.15	194.25
	15	194.10	194.80	195.30	195.70	195.80	195.60	195.35	194.85	194.50	194.10	194.20	194.40
1970	1	194.50	194.90	195.10	195.15	195.15	195.15	195.05	194.75	194.30	193.95	193.90	194.15
	15	194.65	195.05	195.15	195.15	195.20	195.15	194.95	194.55	194.05	193.90	194.05	194.30
1971	1	194.50	194.80	195.25	195.60	196.15	196.40	196.45	196.25	195.90	195.70	195.80	196.10
	15	194.60	195.20	195.35	195.70	196.30	196.55	196.40	196.10	195.90	195.75	196.00	196.25
1972	1	196.55	196.95	197.30	197.50	197.65	197.75	197.65	197.30	196.90	196.60	196.65	196.90
	15	196.70	197.10	197.40	197.65	197.75	197.75	197.50	197.15	196.75	196.60	196.80	196.95
1973	1	197.10	197.45	197.80	198.30	198.50	198.55	198.40	198.15	197.75	197.55	197.40	197.55
	15	197.25	197.60	198.10	198.35	198.50	198.50	198.35	198.00	197.70	197.40	197.45	197.65
1974	1	197.90	198.30	198.60	198.85	199.10	199.10	198.95	198.60	198.10	197.70	197.65	197.70
	15	198.05	198.40	198.80	199.00	199.10	199.05	198.75	198.35	197.90	197.60	197.70	197.75
1975	1	197.75	198.05	198.30	198.50	198.65	198.95	199.00	199.00	198.75	198.60	198.55	198.60
	15	197.90	198.15	198.35	198.55	198.80	199.00	199.05	198.95	198.65	198.55	198.55	198.65
1976	1	198.85	199.15	199.55	199.80	200.10	200.10	199.95	199.60	199.30	199.05	198.85	198.75
	15	199.00	199.35	199.65	200.00	200.10	200.00	199.85	199.45	199.20	199.00	198.85	198.80
1977	1	198.80	198.90	199.05	199.10	199.05	199.20	198.90	198.40	198.15	197.85	197.65	197.60
	15	198.85	199.00	199.05	199.10	199.00	199.10	198.65	198.20	197.95	197.75	197.60	197.60
1978	1	197.65	197.85	198.10	198.35	198.40	198.40	198.20	197.80	197.35	197.25	197.05	197.05
	15	197.70	198.05	198.30	198.40	198.40	198.40	198.05	197.55	197.25	197.15	197.05	197.15
1979	1	197.30	197.55	197.85	198.15	198.15	198.10	197.80	197.35	196.95	196.60	196.40	196.35
	15	[illegible].40	197.65	198.00	198.15	198.20	197.95	197.60	197.20	196.75	196.45	196.40	196.40
1980	1	196.45	196.80	197.30	197.55	197.70	198.15	198.25	198.05	197.75	197.60	197.50	197.60
	15	196.65	196.95	197.55	197.55	197.90	198.30	198.20	197.85	197.65	197.50	197.55	197.65
1981	1	197.85	198.05	198.20	198.35	198.40	198.45	198.20	197.75	197.35	197.00	197.05	197.10
	15	197.90	198.15	198.30	198.40	198.35	198.40	198.05	197.60	197.20	197.15	197.10	197.15
1982	1	197.30	197.40	197.60	197.95	198.20	198.45	198.40	198.35	198.10	198.40	198.55	198.80
	15	197.35	197.50	197.80	198.15	198.40	198.45	198.35	198.20	197.95	198.45	198.60	199.00
1983	1	199.20	199.65	200.00	200.45	200.75	201.15	201.35	201.45	201.60	201.85	202.15	202.50
	15	199.35	199.95	200.20	200.60	201.00	201.35	201.45	201.50	201.75	202.00	202.25	202.75
1984	1	203.10	203.60	204.10	204.45	204.80	205.05	205.35	205.60	206.60	206.90	207.05	207.35
	15	203.40	203.80	204.20	204.60	204.85	205.30	205.40	206.20	206.85	207.00	207.20	207.45
1985	1	207.70	208.00	208.30									
	15	207.90	208.10	208.50									

APPENDIX 2. PARTICIPANT LIST

E. Eric Adams, Department of Civil Engineering, M.I.T., Building 48-311, Cambridge, MA 02139

Ted Arnow, Water Resources Division, USGS, 1745 W. 1700 S., Salt Lake City, UT 84104

Genevieve Atwood, Utah Geological & Mineral Survey, 606 Black Hawk Way, Salt Lake City, UT 84108

Roger G. Barry, CIRES, University of Colorado, Box 449, Boulder, CO 80309

Peter Behrens, Great Salt Lake Minerals & Chemicals Corp., P.O. Box 1190, Ogden, UT 84402

Gail Bingham, Utah State Climatologist, P.O. Box 376, Hyde Park, UT 84318

Raymond J. Bradley, Department of Geology & Geography, University of Massachusetts, Amherst, MA 01003

Raphael Bras, Department of Civil Engineering, M.I.T., Building 48-311, Cambridge, MA 02139

William M. Brown, III, USGS, 345 Middlefield Rd., Mail Stop 998, Menlo Park, CA 94025

David M. Carpenter, National Weather Service, 337 North 2370 West, Salt Lake City, UT 84116

J. R. Chamberlain, Utah Department of Transportation, 4501 S. 2700 W., Salt Lake City, UT 84119

Kathleen Clarke, Congressman Hansen's Office, 324-25th Street, Ogden, UT 84401

Thomas E. Croley, II, Great Lakes Environmental Research Lab., 2300 Washtenaw Ave., Ann Arbor, MI 48104

Donald R. Currey, Department of Geography, University of Utah, Salt Lake City, UT 84112

Daniel Dake, Federal Highway Administration, 125 S. State, P.O. Box 11563, Salt Lake City, UT 84147

Jay Devashrayee, Utah Department of Transportation, 4501 S. 2700 W., Salt Lake City, UT 84119

Henry F. Diaz, NOAA/ERL, 325 Broadway, Boulder, CO 80303

Gregor Eberli, Geologisches Institut, E.T.H., Zurich CH-8092, Switzerland

Stan Elmer, State Lands & Forestry, 355 W. North Temple, 3 Triad, STE 400, Salt Lake City, UT 84010

Don M. Garrett, U.S. Army, Corps of Engineers, 650 Capitol Mall, Sacramento, CA 95814-4794

J. E. Geisler, Department of Meteorology, University of Utah, Salt Lake City, UT 84112

Rick Gore, National Geographic, 1145 17th Street, N.W., Washington, D.C. 20036

Susan A. Green, Department of Geography, University of Utah, Salt Lake City, UT 84112

Myszka Guzkowska, School of Geography, Mansfield RD, Oxford, U.K. OX1 3TB

David E. Hansen, Vaughn Hansen Associates, 5620 S. 1475 E., Salt Lake City, UT 84121

Larry Harrison, Federal Highway Administration, 555 Zang St., Rm 400, Lakewood, CO 80228

Alan D. Hecht, National Climate Program Office, 11400 Rockville Pike, Rm 108, Olney, MD 20832

Herb Hereth, U.S. Army, Corps of Engineers, 650 Capitol Mall, Sacramento, CA 95814

Peter Hovingh, 721 Second Ave., Salt Lake City, UT 84103

L. Douglas James, Utah Water Research Lab., Utah State University, UMC 82, Logan, UT 84322

Thomas R. Karl, National Climatic Data Center, E/CC31, Asheville, NC 28801

Paul A. Kay, Department of Geography, University of Utah, Salt Lake City, UT 84112

Jeff Keaton, Dames & Moore, 250 E. Broadway, Salt Lake City, UT 84111

John W. Lindemann, Consulting Geologist, 845 Emerald Street, Broomfield, CO 80202

Harry F. Lins, USGS, 410 National Center, Reston, VA 22092

David Lovell, Salt Lake County Flood Control & Water Quality, 2033 S. State Street, Salt Lake City, UT 84115

Don R. Mabey, Utah Geological & Mineral Survey, 606 Black Hawk Way, Salt Lake City, UT 84108

David B. Madsen, Utah State Historical Society, 300 Rio Grande, Salt Lake City, UT 84101

W.R. McCormick, Amax Magnesium, 2335 Berkeley Street, Salt Lake City, UT 84109

Judith A. McKenzie, Department of Geology, University of Florida, Gainsville, FL 32611

David M. Meko, Tree-Ring Research Lab., University of Arizona, Tucson, AZ 85721

Randy G. Moon, State of Utah Science Advisor, 116 State Capital Bldg., Salt Lake City, UT 84114

Peter Morrisette, University of Colorado, Campus Box 260, Boulder, CO 80309

Jack Oviatt, Utah Geological & Mineral Survey, 606 Black Hawk Way, Salt Lake City, UT 84108

Clancy Philipsborn, FEMA-Region 8, Building 710, Denver Federal Cntr., Denver, CO 80225

Colin S. Ramage, NOAA/ERL/CIRES, University of Colorado, Campus Box 449, Boulder, CO 80309

Susan Richards, Director Constituent Services, 88 W. 100 N., Rm 105, Provo, UT 84601

Preston Richey, Great Salt Lake Minerals & Chemicals Corp., P.O. Box 1190, Ogden, UT 84402

William E. Riebsame, Natural Hazards Research & Applications Information Center, University of Colorado, Boulder, CO 80309

Richard Roberts, Utah Department of Transportation, 4501 S. 2700 W., Salt Lake City, UT 84119

David Shank, Jr., Geologist, 250 E. Broadway, Salt Lake City, UT 84111

George I. Smith, USGS, 345 Middlefield Rd, MS902, Menlo Park, CA 94025

Normal Stauffer, Utah State Division of Water Resources, 1636 W. North Temple, Salt Lake City, UT 84116

Nancy Stone, FEMA-Region 8, Building 710, Denver Federal Center, Denver, CO 80225

Paul A. Sturm, Utah Geological & Mineral Survey, 606 Black Hawk Way, Salt Lake City, UT 84108

Tom Tripp, Amax Magnesium, 238 N. 2200 W., Salt Lake City, UT 84111

Kidd Waddell, USGS, 1745 W. 1700 S., Salt Lake City, UT 84104

Joseph A. Warburton, Director-Atmospheric Sciences, Desert Research Institute, Reno, NV 89506

Bob L. Whaley, Soil Conservation Service, P.O. Box 11350, Salt Lake City, UT 84147

Gerald Williams, Colorado Basin River Forecast Center, 337 N. 2370 W., Salt Lake City, UT 84112

Doris Wilson, Congressman Nielsen's Office, 125 South State St., Rm 2205, Salt Lake City, UT 84138

Wally Wright, Owner, Saltair Resort, 457 E. 3rd S., Salt Lake City, UT 84111

Pamela J. Young, National Climatic Data Center, E/CC31, Asheville, NC 28801