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Transient modelling of lacustrine regressions: two case studies from the Andean Altiplano

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Abstract:

A model was developed for estimating the delay between a change in climatic conditions and the corresponding fall of water level in large lakes. The input data include: rainfall, temperature, extraterrestrial radiation and astronomical mid-month daylight hours. The model uses two empirical coefficients for computing the potential evaporation and one parameter for the soil capacity. The case studies are two subcatchments of the Altiplano (196 000 km^2), in which the central low points are Lake Titicaca and a salar corresponding to the desiccation of the Tauca palaeolake. During the Holocene, the two catchments experienced a 100 m fall in water level corresponding to a decrease in water surface area of 3586 km^2 and 55 000 km^2, respectively. Under modern climatic conditions with a marked rainy season, the model allows simulation of water levels in good agreement with the observations: 3810 m a.s.l. for Lake Titicaca and lack of permanent wide ponds in the southern subcatchment. Simulations were carried out under different climatic conditions that might explain the Holocene fall in water level. Computed results show quite different behaviour for the two subcatchments. For the northern subcatchment, the time required for the 100 m fall in lake-level ranges between 200 and 2000 years when, compared with the present conditions, (i) the rainfall is decreased by 15% (640 mm/year), or (ii) the temperature is increased by 5·5 °C, or (iii) rainfall is distributed equally over the year. For the southern subcatchment (Tauca palaeolake), the time required for a 100 m decrease in water level ranges between 50 and 100 years. This decrease requires precipitation values lower than 330 mm/year. Copyright © 2004 John Wiley & Sons, Ltd.

KEY WORDS palaeoclimate; lumped model; Quaternary; Peru; Bolivia; lake recession; hydrologic budget

INTRODUCTION

The purpose of this study is to evaluate the delay between climate change and the corresponding water level variations of large lakes. To achieve this objective, a hydrological model was developed for the entire endorheic catchment of the Altiplano. Validated by hydrological observations for the past 25 years, palaeoclimatic reconstructions, i.e. changes in the water input (precipitation) and output (evaporation linked to temperature) are used to explain the vanishing of palaeolake Tauca between 14 000 and 7000 cal. year BP as described by Servant and Fontes (1978) and Sylvestre *et al*. (1999) and the fall in surface level of Lake Titicaca that occurred during the arid period after 11 000 cal. year BP.

The lake-balance model, with 3-month time-steps, calculates the time taken to reach the level of 3707 m a.s.l. in the Titicaca low lake phase and the level of 3656 m a.s.l. for the disappearance of palaeolake Tauca. This model links climate to the hydrology in closed-basin lakes, with two reservoirs: the soil of the watershed and the lake itself. It focuses on the changing ratio between the lake surface area to that of the overall watershed.

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Palaeoclimatological studies in South America: Maslin and Burns (2000) in Amazonia, Thompson *et al*. (1998) in Bolivia and Rech *et al*. (2000) in Chile, show that the intertropical zone underwent great climate, and therefore hydrological, changes during the Quaternary Period. Several studies have recently provided new information on relationships between changes in climate and hydrology in the Central Andes in the northern part of the Altiplano (Abbott *et al*., 1997; Binford *et al*., 1997 Seltzer *et al*., 1998; Talbi *et al*., 1999; Cross *et al*., 2000, 2001; Baker *et al*., 2001a), in its southern part (Baker *et al*., 2001b) and in both areas (Coudrain *et al*., 2002). Argollo and Mourguiart (2000) have given qualitative evidence of climate changes on the Bolivian Altiplano; they successively describe cool-dry conditions at the Last Glacial Maximum (LGM) between 26 and 14 cal. kyears BP followed by a humid episode (14–10·5 kyears BP) and by an arid period between 10·5 and 8 cal. kyears BP.

For the southern Altiplano, quantitative modelling in steady-state conditions shows that to create the palaeolake Tauca (3760 m a.s.l.), the present rainfall rate has to be increased by 48%, with a moderate evaporation rate (Blodgett *et al*., 1996), or increase the current rainfall by 78% with the present evaporation rate (Hastenrath and Kutzbach, 1985).

In this paper, we attempt to provide new climatic values. The originality of our approach in the hydrological balance modelling of the Altiplano is: (i) to calculate the time it takes to reach a certain level; (ii) to consider the changing watershed/lake-surface-area ratio; (iii) to take into account the seasonal distribution in the climatic input data with a quarterly time-step; and (iii) to calculate the potential evaporation on the basis of the rainfall amount.

STUDY SITE AND DATA

Physiographic features

The endorheic catchment of the Peruvian–Bolivian Altiplano includes two subcatchments for which the central low points are Lake Titicaca in the north and Salar de Uyuni in the south (Figure 1). This basin is bounded to the west by the Cordillera Occidental and to the east by the Cordillera Oriental.

The northern subcatchment (14 °S–16·5 °S, 69 °W–73 °W; surface area 56 275 km^2) is a high mountainous basin drained by the outlet of Lake Titicaca to the Rio Desaguadero. The mean water level of the lake is 3810 m a.s.l. with a corresponding surface area of about 8448 km^2. The Lake Titicaca outlet sill lies at 3807 m a.s.l., so that discharge from the lake is zero if the lake level drops below this elevation. To determine the area–volume–elevation relationships for Lake Titicaca we used a data set provided by the 'Plan Director Global Binacional de Proteccion—Prevencion de Inundaciones y aprovechamiento de los recursos del Lago Titicaca, rio Desagvadero, lago Poopo y lago Salar de Coipasa' (TDPS, 1996) (Figure 2).

The southern subcatchment (17 °S–22 °S,66 °W–70 °W) is a high and flat plateau of 140 679 km^2. To determine the area–volume–elevation relationships of the southern subwatershed we used the digital elevation model (DEM) GTOPO30 data set provided by the US Geological Survey EROS Data Center in Sioux Falls, South Dakota. To achieve consistency with the altitude–volume–surface area deduced from the digitized 1 : 50 000 maps in Blodgett *et al*. (1996), it was necessary to decrease the elevation by 28 m from GTOPO30. The correlation coefficient (R^2) between these two independent DEMs is around 0·98 for the surface-area–elevation relationship and 0·99 for the volume–elevation relationship (see Figure 2). With this revised DEM, we established area–volume–altitude relationships for the palaeolake in the southern subcatchment that can be used over the entire altitude range of the watershed.

The two subcatchments have morphological differences (Figure 2). For the same modification in water level, the water volume and the free water surface are one order of magnitude higher in the southern flat catchment than in the northern one.

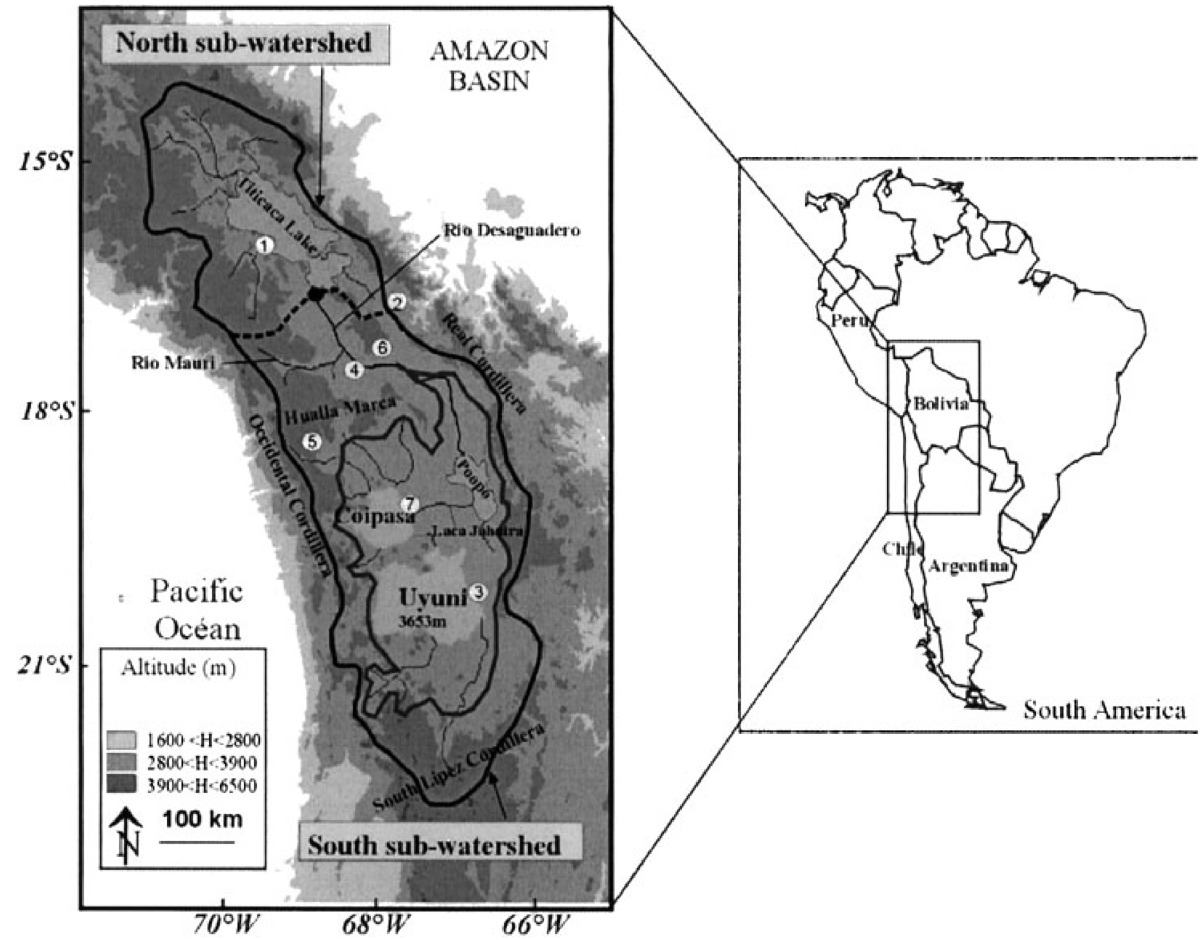

Figure 1. Location of the study area: bold solid line marks the boundary of the endorheic watershed (197 000 km^2); dashed line marks the internal boundary between the northern subwatershed (56 275 km^2) including Lake Titicaca and the southern subwatershed (140 679 km^2) with Lake Poopo and the salars of Coipasa and Uyuni. The only hydraulic connection between the two subwatersheds is the rio Desaguadero (filled circle). The maximum surface area of the palaeolake Tauca is shown by a grey solid line (55 000 km^2). Circled numbers are the climatological stations: 1, Puno; 2, LaPaz; 3, Uyuni; 4, Ulloma; 5, Sajama; 6, Patacamaya; 7, Coipasa

Modern hydroclimatic features

There is a negative rainfall gradient from north (about 750 mm/year) to south (about 160 mm/year at Uyuni) with a marked rainy season (60% of the annual amount during the 3 months between December and February, DJF) and a long dry season (March to November). During the austral summer, the precipitation is a consequence of the movement of the intertropical convergence zone (ITCZ) (Ronchail *et al*. 1995; Garreaud and Aceituno, 2000) and corresponds to intense convective rainstorms when the east winds in the middle and upper troposphere permit the advection of moisture from the Amazon basin. The remaining 40% of precipitation during the austral winter corresponds to the northward movement of cold fronts (May and June) and to 'cut-offs' (July and August), as described by Vuille and Baumgartner (1998) and Vuille and Ammann (1997). Seasonality of the precipitation depends on the relative importance of these processes, i.e. the ITCZ movement and the movement of cold fronts and occurrence of cut-offs.

In the northern subcatchment, an estimate of the potential evaporation rate is 1600 mm/year after a review of the evaporation of Lake Titicaca (Pouyaud, 1993). From measurements at climatological stations at Puno

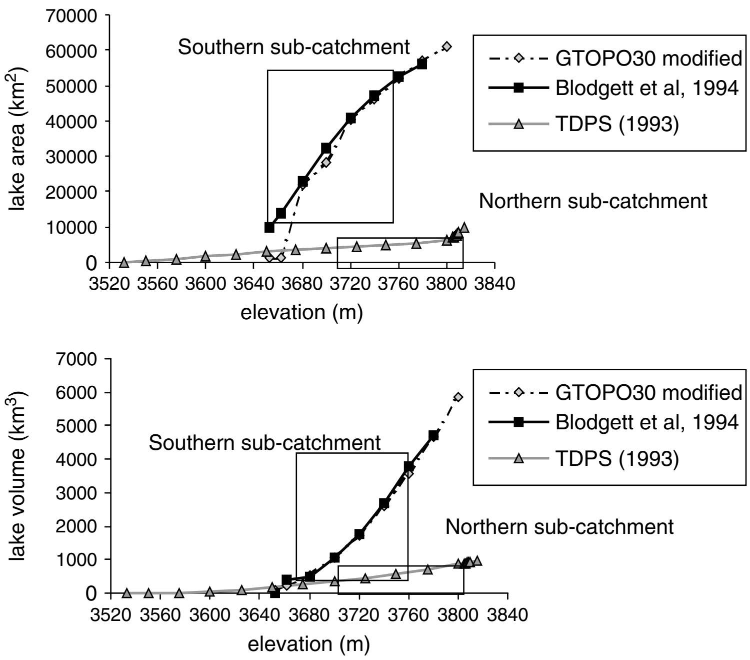

Figure 2. Topographic relationships for the northern and southern subwatersheds. For the southern one, the two relationships are the shifted digital terrain model GTOPO-30 DEM and the digitized topographic maps from Blodgett's study (1996). The rectangles indicate the scale range of changes simulated in the present paper

and La Paz (Figure 1) and from hydrological studies reported in Roche *et al*. (1992) and TDPS (1993) the average precipitation for 1960–1990 was 750 mm/year and the average discharge of the Rio Desaguadero at the outlet of Lake Titicaca was 1 km^3/year (32 m^3/s).

For the southern subcatchment, data are available from Uyuni, Ulloma, Sajama, Patacamaya and Coipasa (Figure 1). The average potential evaporation and precipitation amounts in the 1960–1990 period were 1600 mm/year and 350 mm/year, respectively (Roche *et al*., 1990).

Past hydroclimatic features

A recent study (Cross *et al*., 2000) based on the analysis of cores taken in Lake Titicaca define a high level, around 3810 m a.s.l. at 10 830 cal. years BP, and a low level, 3710 m a.s.l. at 6490 cal years BP. The surface areas associated with these two lake levels are 8448 km^2 and 4862 km^2 (as reported in Talbi *et al*., 1999).

The evolution of the water level in the southern catchment is based on the elevation and age determination of palaeobioherms precisely described in Rouchy *et al*. (1996). The palaeobioherms were built up by different plant assemblages composed of green algal microflora associated with cyanobacterial communities. Their morphology is similar to those of stromatolites and their altitude is linked to the level of the palaeolake. The highest altitude of the Tauca water level was first assumed to be 3720 m (Servant and Fontes, 1978). Higher altitude's have been indicated by different authors (Rouchy *et al*., 1996; Sylvestre *et al*., 1999). In order to improve our knowledge of the highest level of the paleolake Tauca, we measured the elevation of the highest terrace of the palaeobioherms. These measurements were made with two Trimble 4000 SSI L1/L2 code GPS receivers. We used a set of five benchmarks set by the South America—Nazca Plate Motion Project (SNAPP) UNAVCO (Norabuena *et al*., 1998). We created temporary bases with baselines of 50–200 km. Base data were collected by a static method over 8–12 h. Data were collected with a baseline length that never exceeded 50 km and the collection time ranged from 20 min to 1·5 h depending on the baseline length. For post-processing we used the L1/L2 code, the WGS84 geodetic system, the EGM96 geoid model and precise ephemeris. The results of the field studies show the highest bioherm palaeoterrace elevation to be between $3751 \pm 0{\cdot}5$ m and $3769 \pm 0{\cdot}5$ m a.s.l. depending on the location. With an uncertainty of ± 10 m, the altitude of the high Tauca terrace can be estimested as 3760 m. The maximum surface area of Lake Tauca was then calculated with the modified digital elevation model GTOPO 30 (see above). The result was $55\,000 \pm 6000$ km^2 (Figure 1).

The hydrological changes that will be taken into account in the following section are a decrease of 100 m that occurred before or around 11 000 years BP in the Titicaca lake and in the southern subcatchment. This

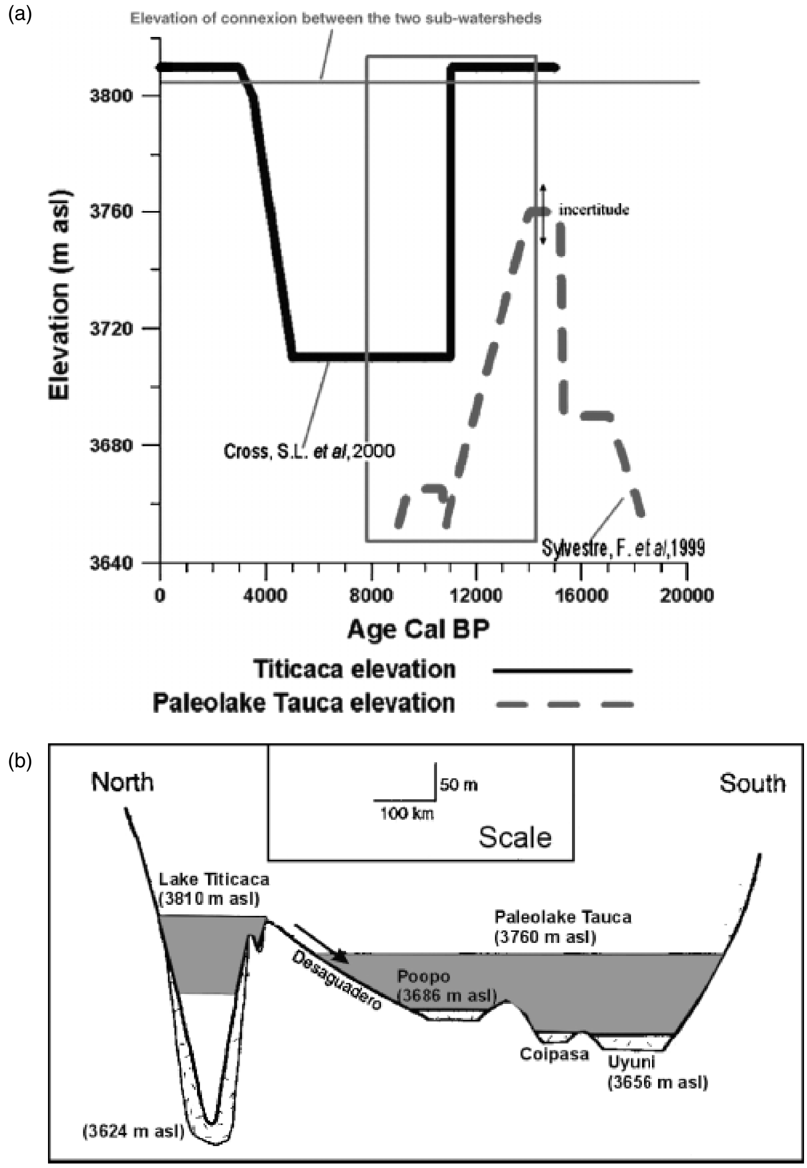

Figure 3. (a) Palaeolake variations during the past 20 000 year for the northern subwatershed (Titicaca), and the southern subwatershed (Tauca). The rectangle limits the studied events with the lacustrine recessions for each subwatershed. The recession time for Lake Titicaca is shorter (around 100 years) than that of Lake Tauca (around 3000 years). (b) Cross-section of the Central Altiplano with volumetric changes (grey zone), adapted from Argollo and Mourguiart (2000)

decline corresponds to a water volume of 1000 km^3 for the northern catchment and of 4500 km^2 for the southern one (Figures 2 and 3).

WATER BALANCE MODEL

Model principle

We developed a reservoir model with 3-month time-steps for input data and 6-day time-steps for the calculations. The model takes into account two reservoirs: the soil (or watershed) and the lake, the surface area of which may change at each time-step depending on the water balance in the two reservoirs. The reservoirs are connected, as the soil recharges the lake. Figure 4 shows the model structure. The input data are precipitation, P_{ws} (mm/quarter) and actual evapotranspiration Et_{ws} (mm/quarter) for the watershed (or soil) reservoir. For the lake reservoir, the input data are precipitation P_{lake} (mm/quarter) and evaporation E_{lake} (mm/quarter). The Desaguadero mean quarter discharge (Q_{des}, in m^3/s) is the northern subcatchment outlet from Lake Titicaca that contributes to the inflow of the southern sub-catchment.

The hydrological balance equation used in the model is

$$\frac{dV_{lake}}{dt} = (P_{lake} - E_{lake})S_{lake} + (P_{ws} - Et_{ws})S_{ws} + \beta Q_{des} \quad (1)$$

with V_{lake} the volume of water of the lake, t the time and $\beta = 1$ for the southern subcatchment and -1 for the northern subcatchment.

At each time-step, the model calculates the volume of the lake. The Deasaguadero outflow, Q_{des}, from the northern lake is computed in the model according to the empirical relationship between measured Q_{des} and water levels of the Titicaca lake (Figure 5). The level of the water and the surface area are then computed using the morphological relationships illustrated in Figure 2. This model was constructed with the VENSIM software package (VENTANA, Inc.), which solves the equations by an automatic fourth-order Runge–Kutta numerical scheme.

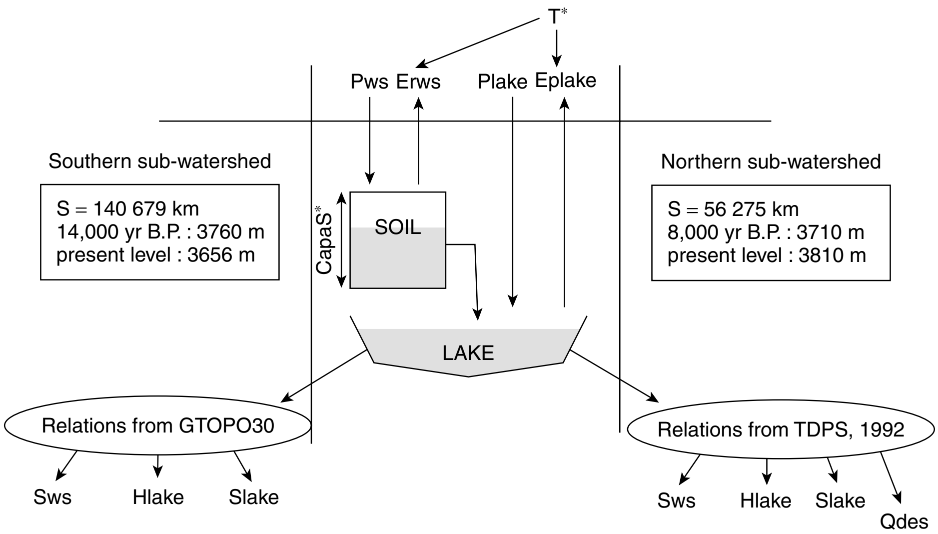

Figure 4. Lumped model diagram; watershed precipitation (Pws), lake precipitation (Plake), watershed evapotranspiration (Erws), lake evaporation (Eplake), watershed surface area (Sws), lake surface area (Slake), lake elevation (Hlake), Desaguadero discharge (Qdes) and soil capacity coefficient (CapaS)

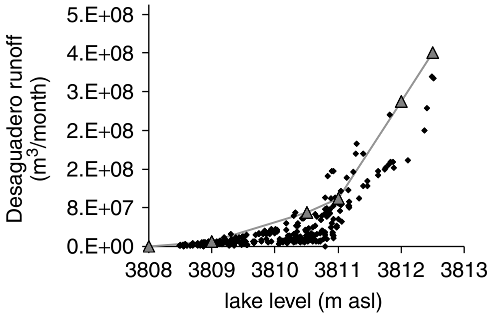

Figure 5. Gauged empirical relationship between the northern lake level (Hlake) and Desaguadero outflow (Qdes). Black dots are empirical points and grey triangles form the envelope curve that serves in the model

Evaporation calculation

Potential evaporation. The choice of a potential-evaporation formula is crucial in a balance-model approach. There are many formula classes to calculate the evaporation and they are generally specific to one set of climate conditions (Xu and Singh, 2000). Vacher *et al.* (1994) made a complete analysis of the potential evaporation in the Bolivian Altiplano, which demonstrated that the most appropriate evaporation formula is based on the energy balance and depends on five terms: (i) net radiation, (ii) overall solar radiation, (iii) albedo, (iv) terrestrial radiation and (v) atmospheric radiation. In our paleoclimatological study, the parsimony of climate parameters is important and furthermore, in a water-balance model, input data complexity is not synonymous with better performance (Hervieu, 2001). We chose the generalized equation of Xu and Singh (2000), which is a radiation based method. This formula (Equation 2) is adapted for a monthly series and depends on two climate variables, only the mid-month air temperature (T in °C) and the total solar radiation (Rg in J/cm^2/day)

$$EP = \frac{Rg}{\lambda}(T + 17{\cdot}8)0{\cdot}0145 \tag{2}$$

$$\lambda = 595 - 0{\cdot}51T \tag{3}$$

where EP is in mm/day, Rg is the total solar radiation (in J/cm^2/day), T the mid-month air temperature (°C) and λ the latent heat (Cal/g) is calculated according to Equation (3).

To compute the total solar radiation, Rg, we used climate data from Puno to establish an empirical relationship based on extraterrestrial radiation Re (J/cm^2/day), on mid-month astronomical daylight hours Sun_th (h/day) and on actual mid-month daylight hours Sun_tr (h/day) as described in Equation (4)

$$Rg = \left(0{\cdot}1 + 0{\cdot}7\frac{\text{Sun\_tr}}{\text{Sun\_th}}\right) Re \tag{4}$$

where Re is the extraterrestrial radiation (J/cm^2/day), Sun_th is the mid-month astronomical daylight hours (h/day) and Sun_tr is the actual mid-month daylight hours (h/day). Finally, the relationship for potential evaporation of the Altiplano can be written

$$EP = \left(0{\cdot}1 + 0{\cdot}7\frac{\text{Sun\_tr}}{\text{Sun\_th}}\right) Re\frac{(T + 17{\cdot}8) \times 0{\cdot}0145}{(595 - 0{\cdot}51T)} \tag{5}$$

At each time-step the evaporation from the lake is computed by Equation (5), and the evapotranspiration from the watershed (or soil) reservoir is computed using a combination of the estimation of the amount of water available and of this equation (see later). The modern annual value calculated with this relationship gives

1670 mm/year and is comparable with the Titicaca lake evaporation value of 1650 ± 100 mm/year given by Talbi *et al*. (1999).

Evapotranspiration and soil module. The actual evapotranspiration is computed from the available water quantity in the soil. The computation is based on the use of a soil reservoir as classically defined in the lumped models GR (Génie Rural) developed by the CEMAGREF, i.e. the French public agricultural and environmental research institute (Makhlouf and Michel, 1994). At each time-step (quarter) this reservoir permits, via a soil parameter 'CapaS' (soil capacity in m of water), the calculation of the soil-water content and of the water available for evapotranspiration (*ETws* in m)

$$Etws_{n+1} = \left(\frac{Hws_n}{\text{CapaS}}\right) \times \left(2 - \frac{Hws_n}{\text{CapaS}}\right) \times EP_{n+1} \tag{6}$$

$$Hws_{n+1} = Hws_n + Pws_{n+1} - Etws_{n+1} \tag{7}$$

where *Hws* is the water content of the soil reservoir and *Pws* is watershed precipitation.

As a conclusion, this conceptual lumped model is very parsimonious, with only one CapaS parameter and two reservoirs (the soil and the lake).

Parameterizations and calibration

To understand the climatic mechanisms we analysed how the input data are linked together. We did this for the northern and the southern subcatchments with the following input data: (i) mid-month daylight hours; (ii) precipitation and (iii) temperature. The correlation coefficient (R^2) between variables shows that daylight duration and precipitation are closely linked. Indeed, high precipitation rates are associated with a thick cloud cover and therefore with weak daylight. To limit the number of input data for the final palaeohydrological calculations, we calculated an empirical relationhip valid for the northern and southern subcatchments with current data over the past 30 years for the seven meteorological stations (see Figure 1 for locations). This linear relationship can be written

$$\text{Sun\_tr} = A \times P + B \tag{8}$$

where Sun_tr is mid-month actual daylight hours (h/day), P is precipitation (mm/quarter), and A and B are empirical parameters calculated from data obtained at the meteorological stations with respective values of $-0{\cdot}008$ and $9{\cdot}18$ for the two subwatersheds ($R^2 = 0{\cdot}8$).

Finally the model is able to run with two astronomical inputs, mid-month theoretical daylight hours (Sun_th) and extraterrestrial radiation (*Re*), two input values, precipitation (P) and temperature (T) and one parameter, the soil capacity (CapaS).

RESULTS

In order to validate the model, we simulated the hydrological behaviour of the northern and southern subcatchments. The calibration model and the sensitivity tests were performed for the recent period for which instrumental hydroclimatological data are available.

Modern simulations

For the Titicaca watershed (northern subcatchment) we chose the historic record from 1960 to 1990, where there are no data gaps and the flow regime is natural for the Desaguadero discharge at the Titicaca outlet. The selected score function is the comparison between the measured and calculated Desaguadero flow at the Titicaca outlet. The flow data were collected at the gauging station of 'Puente Internacional' (69°02'W, 16°33'S) (Figure 1). The calibration routine is the linear Powell optimization. The results are shown in Figure 6a and the associated lake-level in Figure 6b.

For CapaS of 0·24 m, the correlation coefficient is $R^2 = 0{\cdot}87$ and we observe a good agreement when the discharge is high during the last decade of the calibration period. The failure to reproduce low flow could be explained by the relationhip that links the lake volume to the Desaguadero flow (see figure 5), as this relationship is uncertain.

The accuracy of the simulations in volume terms also can be evaluated with the balance criterion (Cr_{bal}) defined in Perrin (2000) where i is the quarter:

$$Cr_{\text{bal}} = 1 - \left| \sqrt{\frac{\sum_{i=1}^{n} Q_{\text{cal},i}}{\sum_{i=1}^{n} Q_{\text{obs},i}}} - \sqrt{\frac{\sum_{i=1}^{n} Q_{\text{obs},i}}{\sum_{i=1}^{n} Q_{\text{calc},i}}} \right| \tag{9}$$

This coefficient is 0·9 for the calibrated CapaS. The sensitivity analysis shows that to keep the correlation coefficient R^2 above 0·8, CapaS should range between 0·2 ($R^2 = 0{\cdot}80$, $Cr_{\text{bal}} = 0{\cdot}60$) and 0·3 ($R^2 = 0{\cdot}84$, $Cr_{\text{bal}} = 0{\cdot}60$).

For the southern subcatchment, we simulated the lake-level variations with equivalent independent records during the 1979–1990 period for the CapaS determined on the northern subcatchment (value between 0·2 and 0·3). Figure 6c shows that the lake level varies by around 0·4 m. This change corresponds to a volume variation of $4{\cdot}7 \times 10^{10}$ m^3. This volume variation is partly explained by that of Lake Poopo during this period, which increased its volume by 4×10^9 m^3 (Taborga, 1994).

Palaeoclimatic reconstructions

For the northern subcatchment, the objective was to simulate a 100 m decrease in the Lake Titicaca level from its initial 3810 m a.s.l. value. To reach this goal, a 1000-year data input series of mean quarterly values was created by using, on the one hand, the equivalent independent record of the 1965–1990 period (see above) for precipitation and temperature and the astronomical daylight hours and the extraterrestrial radiation data at latitude $-16°$ obtained from Berger and Loutre (personal commmunication) calculated for the beginning of the Holocene at 10 000 years BP. Based on the hypothetical 1000-year data series, climate scenarios were then tested according to: (i) an increase in air temperature; (ii) a decrease in precipitation; (iii) presence or absence of the seasonal distribution of rainfall related to the CFCOP.

Another assumption is to consider the linear relationship between daylight hours and precipitation established under modern conditions unchanged during the Pleistocene to Holocene period. This assumption could be adopted considering that the relationship is valid from the northern to southern Altiplano where different climatic conditions occur. Furthermore, the soil capacity, CapaS, is taken from modern value (between 0·2 and 0·3) because at that time no drastic changes in the geology and in the vegetation cover occurs (Markgraf, 1989). When the precipitation rates are changed from the present values without changes in the distribution over the year, a 15% decrease from the present P value (640 mm/year) is the minimum decrease that reduces the level of Lake Titicaca by 100 m in 700 years with the calibrated CapaS (Figure 7a). The tested precipitation range was 300 to 750 mm/year.

As shown in Figure 7b, a rise in air temperature has a strong impact: when we increased the temperature by 5·5 °C, while keeping the present rainfall conditions, we obtained a lake-level decrease of 100 m in 850 years owing to the rise in the potential evaporation.

The impact of the seasonal distribution is also significant (Figure 7c). The scenario with the same precipitation for each quarter shows that only 200 years are needed to draw down lake level by 100 m with a precipitation decrease of 15%. The low lake level could even be reached in 1000 years with modern

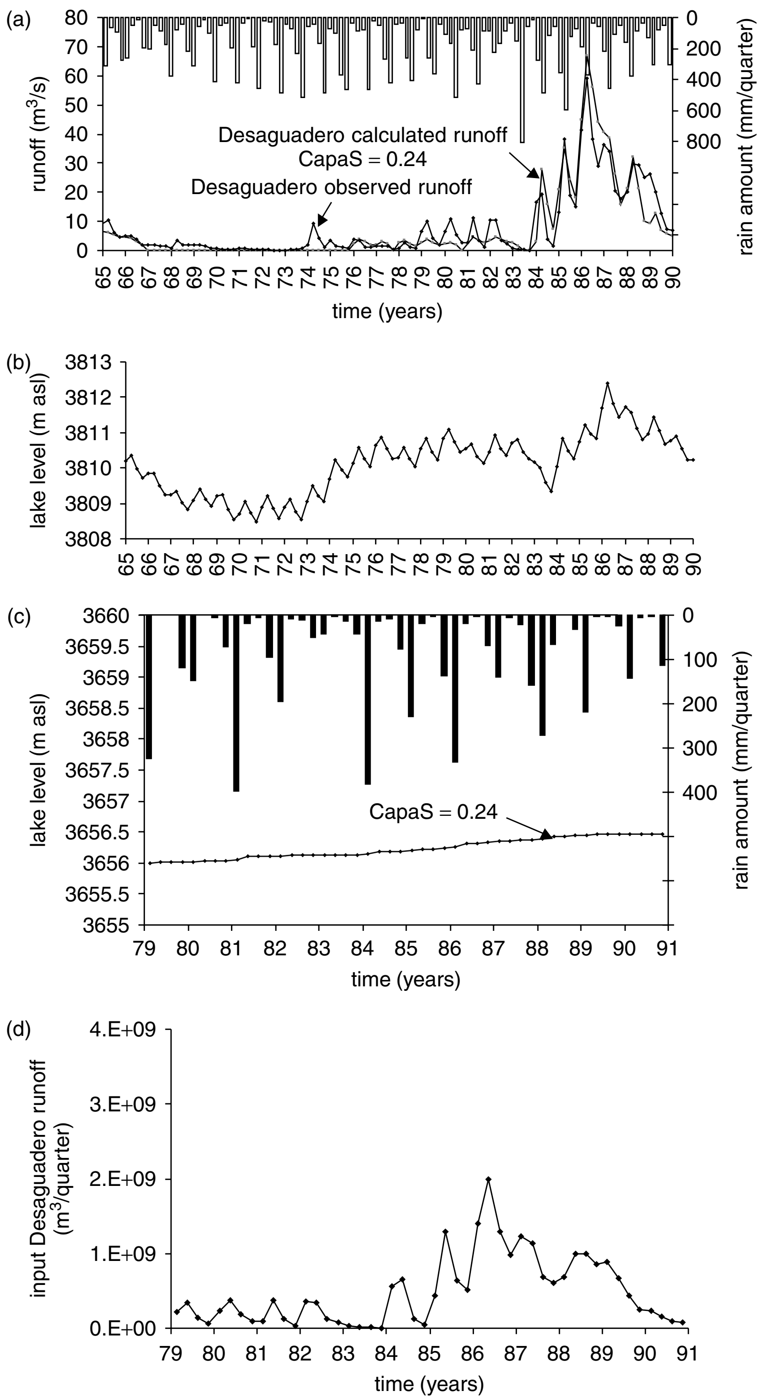

Figure 6. (a) Northern subcatchment simulation results for the Desaquadero discharge at quarterly time-steps (with seasonal distribution), 1965–1990. (b) Northern subcatchment simulation results for the Titicaca lake-level variation at quarterly time-steps (with seasonal distribution), 1965–1990. (c) Southern subcatchment simulation results for lake-level variations at quarterly time-steps (with seasonal distribution), 1979–1990 period. (d) Southern subcatchment input Desaguadero runoff at quarterly time-steps (with seasonal distribution), 1979–1990 period

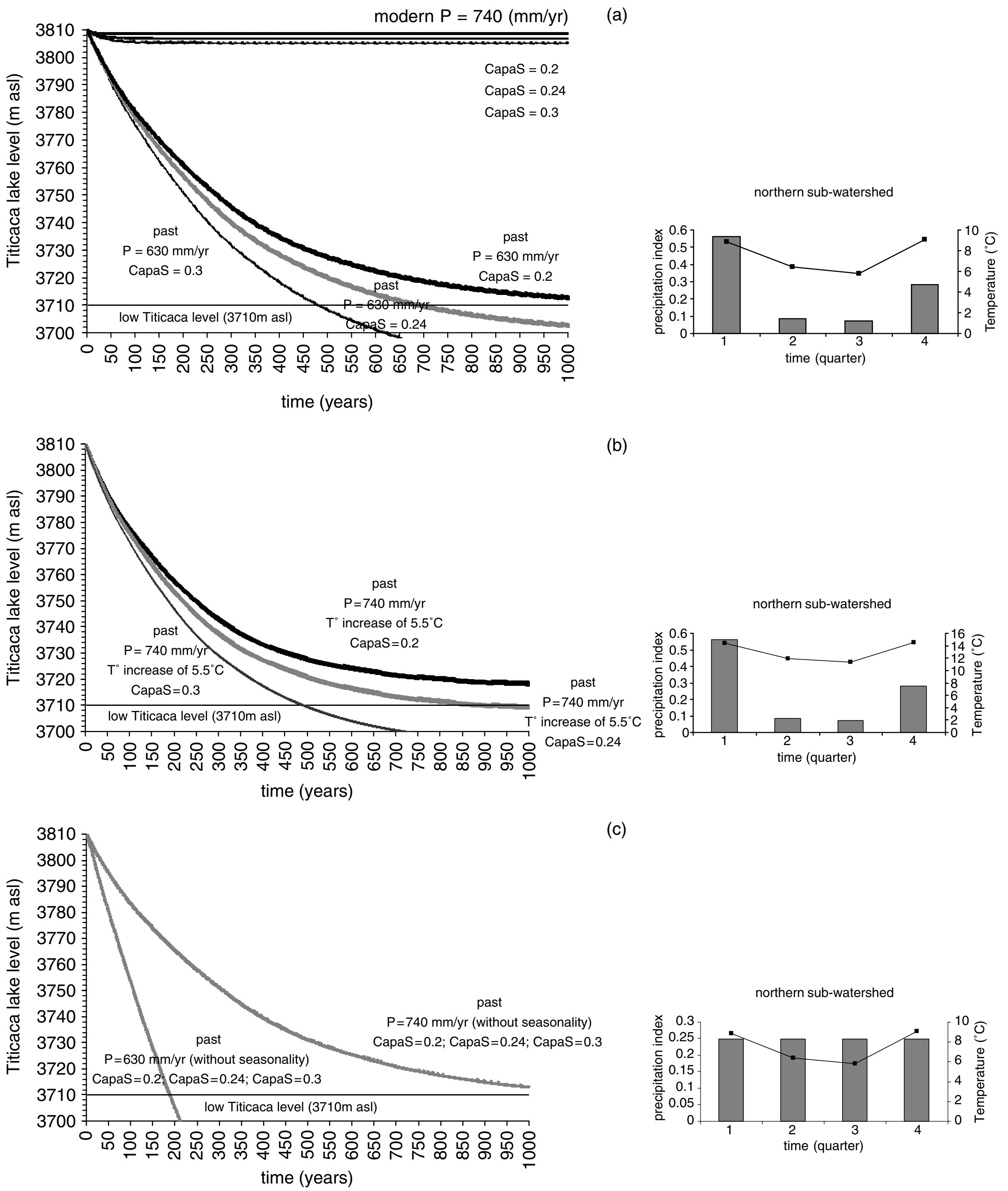

Figure 7. For each simulation, we indicate at the right-hand side the annual distribution of precipitation (with or without seasonal distribution) and annual distribution of temperature used. (a) Northern subcatchment simulation results for quarterly time-steps, 1000 simulation years, influence of the amount of rainfall. (b) Northern subcatchment simulation results for quarterly time-steps, 1000 simulation years, influence of temperature. (c) Northern subcatchment simulation results for quarterly time-steps, 1000 simulation years, influence of seasonal distribution

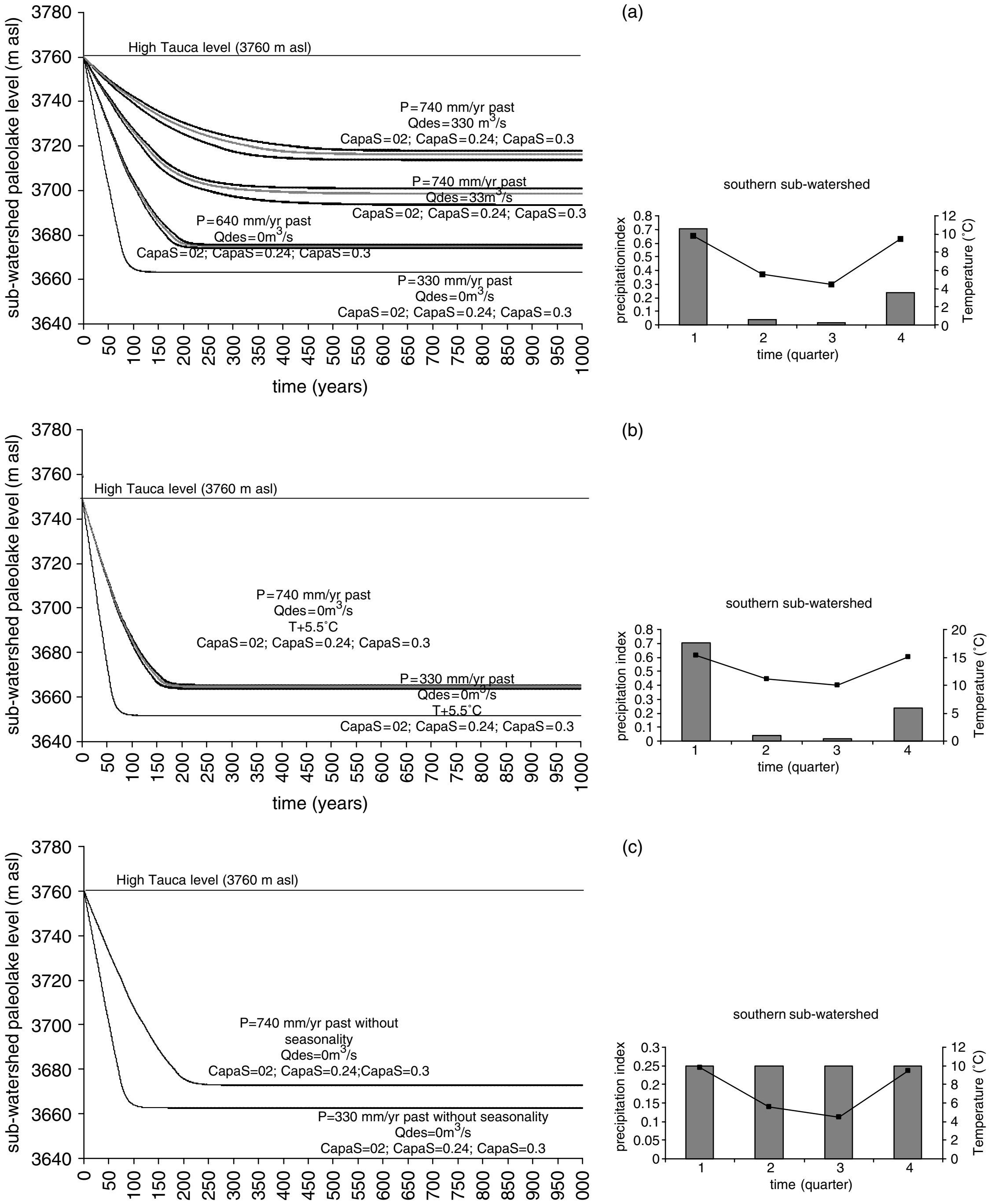

Figure 8. For each simulation we indicate at the right-hand side the annual distribution of precipitation (with or without seasonal distribution) and annual distribution of temperature used. (a) Southern subcatchment simulation results for quarterly time-steps, 1000 simulation years, influence of rain-fall amount and of the Desaguadero outflow. (b) Southern subcatchment simulation results for quarterly time-steps, 1000 simulation years, influence of temperature. (c) Southern subcatchment simulation results for quarterly time-steps, 1000 simulation years, influence of seasonal distribution

precipitation equally distributed over the four quarters. In the case without seasonal precipitation, the soil-water content can respond to the potential evaporation demand throughout the year. In the case with a marked wet-season, the recharge to the lake is higher. Sensitivity tests on the precipitation-decrease scenario show that the lake level is very sensitive to the CapaS parameter (Figure 7a).

For the southern subwatershed, we simulated the Lake Tauca decrease from 3760 to 3656 m a.s.l. As for the northern subwatershed, a 1000-year input data-series with mean values for each quarter was created. The climatological input was calculated with the present values from the southern subcatchment and astronomical data for the latitude 20° south were provided by Berger and Loutre (personal commmunication). The simulations show that the stable Tauca high-water level is obtained with an annual precipitation of 900 mm. From this high level, with present Desaguadero outflow values multiplied by 10 (330 m^3/s), the lake decreases to 3718 m a.s.l. (Figure 8a). With Desaguadero outflow equal to the modern value and with a precipitation value of 740 mm/year, the lake level decreases to 3700 m a.s.l. Finally, to achieve the low level, precipitation must be lower than 330 mm/year (Figure 8a).

Figures 8b and 8c demonstrate that changes in temperature or rainfall distribution over the year would not radically change the time it takes to reach the low level, which is around 100 years. Therefore, if the true decrease time was longer, the climate changes cannot be instantaneous.

DISCUSSION AND CONCLUSION

This study deals with the impacts of climate change on the past and present hydrological conditions throughout the central Altiplano. We demonstrate that a quarterly lumped model can simulate the present hydrological conditions where Lake Titicaca is recharged by runoff. For the simulations of the past up to the arid period (after 10 000 years BP), the following conclusions can be drawn: for the northern subcatchment a 15% decrease of the modern precipitation is sufficient to explain the fall by 100 m of Lake Titicaca in 700 years; the effects of temperature demonstrate that an increase of 5·5 °C from the present (with precipitation amounts at modern values) permits this decrease in 850 years. The disappearance of the seasonal distribution of precipitation (with a decrease of 15% from modern amounts) accelerates the lowering of the lake-level, which is reached in 200 years. For the southern subwatershed, 100 years are necessary to decrease the lake level by 100 m when the precipitation amount is below 300 mm/year.

The impact of the seasonal rainfall distribution is clearly demonstrated and justifies the selected time-step (3 months) of the model. Although one climate change scenario cannot be singled out from our calculations, at least our model can test the plausibility of any proposed scenario, and give bounds to the magnitude of the changes in precipitation, temperature and seasonal distribution required to reproduce the past inferred lake-level changes.

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