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On the Ice Age Glaciation of the Tibetan Highlands and its Transformation into a 3-D Model

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ABSTRACT: We present an interdisciplinary study on data and modeling intercomparison, concerning the possible existence of a Tibetan ice sheet and its climatological implications during the ice age. In the ice sheet model the fields of ice flow and temperature are calculated, and a highly parameterized formulation of the yearly snow balance is used, defining the forcing at the surface of the ice sheet. The data set used, supplies the height of the equilibrium line of the glaciers (=ELA) and documents the maximum extension of the glaciated areas. With prescribed snow accumulation above the ELA and melting below, the model is integrated for 10 000 model years and the model glaciation is then compared with the data.

The main results are: Provided the height of the glacial equilibrium line has been reconstructed correctly, a Tibetan ice sheet can be bult up within 10 000 model years, using moderate rates of precipitation (maximum snow fall: 100 mm/year). Comparison of data and model glaciation suggests an increase of precipitation from the NW to the E of Tibet and from the S to the NE, which reflects the presently observed pattern of the monsoon circulation.

Introduction

During the last 10–20 years, an increasingly detailed data base has been collected on the paleo climate. These proxy data have been extracted from sediments on land, the deep sea and from ice cores. In general, the data have to be transformed into climatological units and the sediment has to be dated before time series of the past climate can be reconstructed. The data base is especially dense for the last 100 000 years (the last ice age cycle). This initiated the development of a hierarchy of climate models, designed to simulate the long time scales of the paleo climate. The central prognostic model component is the ice sheet, while the two other major components of the climate system (the atmosphere and the ocean) follow the boundary conditions, partly set by the ice sheet, in a quasi-stationary way. In turn, the coupled atmosphere-ocean system determines the yearly glacier equilibrium line altitude (ELA), which is the most important upper boundary condition for the ice sheet. With simple 2-d ice sheet models (North-South profile) and a highly parameterized form of the snow balance, driving the ice sheet model, some characteristic features of the changes in the global ice volume can already be simulated (see Oerlemans and van der Veen 1984, for a review). But with the recent development of 3-d ice sheet models (Jenssen 1977; Herterich 1988) and the use of last-generation computer facilities, we are now also able to utilize more detailed information, contained in the geologic record, such as the glaciated area and the height of the glacial equilibrium line. Such data-model comparison studies have an interdisciplinary character. Both the interpretation of detailed data sets and running complex numerical models afford some specific experience. In this work, we would like to demonstrate the merits of such an interdisciplinary research effort by applying this method to the question of the possible existence of a Tibetan ice sheet during the ice age. The data set which provides information on the height of the glacial equilibrium line and on the maximum extent of the glaciated area is described in section 2. In section 3 the ice sheet model is presented, followed by modeling results concerning the build-up of the Tibetan ice sheet. In the final section 4 we will discuss these results in the context of its implications for modeling the glacial cycles.

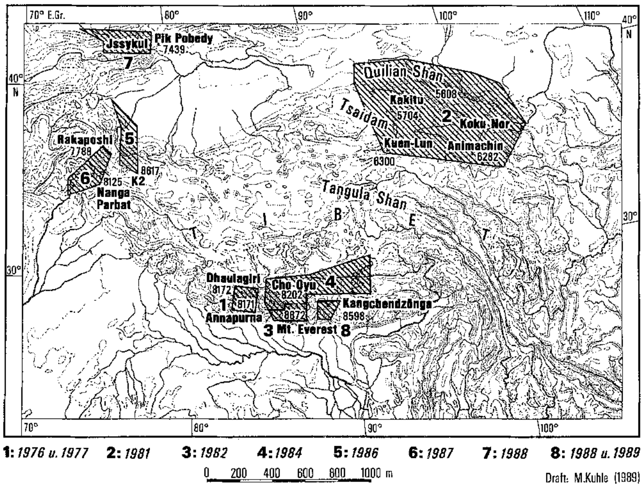

Fig 1 Research areas in High Asia visited in the course of seven expeditions

Data

The Possible Size of a Pleistocene Tibetan Ice Sheet

The aim of nine expeditions (between 1976 and 1989) was to reconstruct the ice age-glaciation in Tibet and in its surrounding mountain areas (Fig 1). Since the results were already published in detail (Kuhle 1980–1988), we will summarize only the most important data to synthesize the glaciated area (Fig 2). At the southern margin of the Tibetan plateau, outlet glaciers penetrated the Himalayas. Moraines could be identified in the lower Mayangdi- and Thak Khola of the Dhaulagiri area (Fig 1, No 1) down to 1100 m asl. 200–300 km to the E, another outlet glacier was flowing through the Bo Chu valley down to 1600 m. The Dudh Kosi glacier, draining the Cho Oyu-, Mt. Everest- and Lhotse-area reached down to 1800 m. All these glaciers were fed by a S Tibetan network of ice streams (Fig 1, No 3 and 4; Fig 2, I3). A South Tibetan glaciation is not only suggested indirectly by the existence of former outlet glaciers but also by direct geologic evidence in areas N and S of the Tsangpo valley: at the Chalamba La pass (29° 41' N/90° 15' E; Fig 2, I2) in 5300 m height blocks of erratic granite lying on rhyolytic rock were found, indicating transport from remote sites. These blocks are placed 200 m above the pass level. Since the bottoms of the valleys to both sides are lying 800 m and 1000 m lower, the local ice sheet thickness should have been at least 1200 m. The ice sheet was probably much thicker, covering the surrounding areas with exception of a few more than 6000 m high mountains. These sharply peaked mountains contrast with the otherwise smoothly shaped, glaciogenic topography. The Lulu valley in the Latzu mountains is another (classic) place indicating a fossil ice sheet (28° 53'N/87° 27' E; Fig 1, No 4; Fig 2, I3). It is carved into basaltic rock and filled with ground moraines in the height range from 4400 m to 5000 m, containing also blocks of two-mica granite. Furthermore, these huge blocks, which were transported from the North, are deposited at least 170 m above the bottom of the valley.

Glaciers from the Nanga Parbat and the Karakoram at the S-W margin of Tibet (Fig 1, No 6) were the source for a network of ice streams, filling the Indus valley system. One of the outlet glaciers reached down to 980 m (Fig 2, Karakoram). Its moraines were dated to be remnants of the Würm ice age (Schroder; Saqid Khan 1988).

In the N-W of Tibet, ice age glaciers spreaded from the Karakoram main ridge, the Aghil- and Kuen-Lun-mountains into the Tarim depression down to 2000 m height (Fig 1, No 5), forming cold Piedmont-type glacier tongues and accumulating 400–600 m thick moraines (Fig 2, behind K2). The long valleys on both sides of the Karakoram main ridge were filled with outlet glaciers originating from Western Tibet. This could be proved by glacier carvings, glaciated valley flanks and erratics which were deposited far above the valley bottom.

Both the N part of the ice sheet I2 (Fig 2) and the margin of ice sheet I1 (Fig 2) to the Tsaidam depression could be documented by a total of 43 moraine deposits in the Animachin, the Kuen Lun and in the Quilian Shan. Using C^{14} -dating, the maximum Tibetan glaciation is older than 8600 to 12800 years BP (Kuhle 1982, 1983; Yamanaka 1982; and geomorphological interpretation Kuhle 1987a, 1988b). C^{14}-dating of lake sediments at about 100 positions on the Tibetan highlands yielded ages not older than 13000 years BP (personal communication with Prof. Xü Daoming, Lanzhou Institute of Glaciology). These measurements suggest a Tibetan glaciation still present during late glacial stages. All these findings – including those of previous research (v. Loczy 1893; Odell 1925; Norin 1932; Trinkler 1932; Weng & Lee 1946 et al.) – point to the existence of a Tibetan ice sheet, covering about $2.4 \cdot 10^6$ km^2.

The Height of the Glacier Equilibrium Line (ELA)

A rough estimate on the depression of the equilibrium line during the ice age can be obtained by dividing the change in the height of a glacier tongue (ice age height minus present height) by a factor 2. A better estimate can be derived, if the height of the catchment area together with the geometry of the glacier bed is taken into account (Kuhle 1986). In all those mountain areas indicated in Fig 1, a lowering of the equilibrium line between 1100 and 1500 m with respect to its present height could be determined. As a consequence, the ice age equilibrium line was up to 600 m below the mean height of the Tibetan plateau (Kuhle 1988a, Fig 2; 1988b). This is an independent argument (besides erratics) in favour of a wide spread glaciation of Central Tibet during the ice age.

The height of the equilibrium line is a function of the local radiation budget, the temperature and precipita-

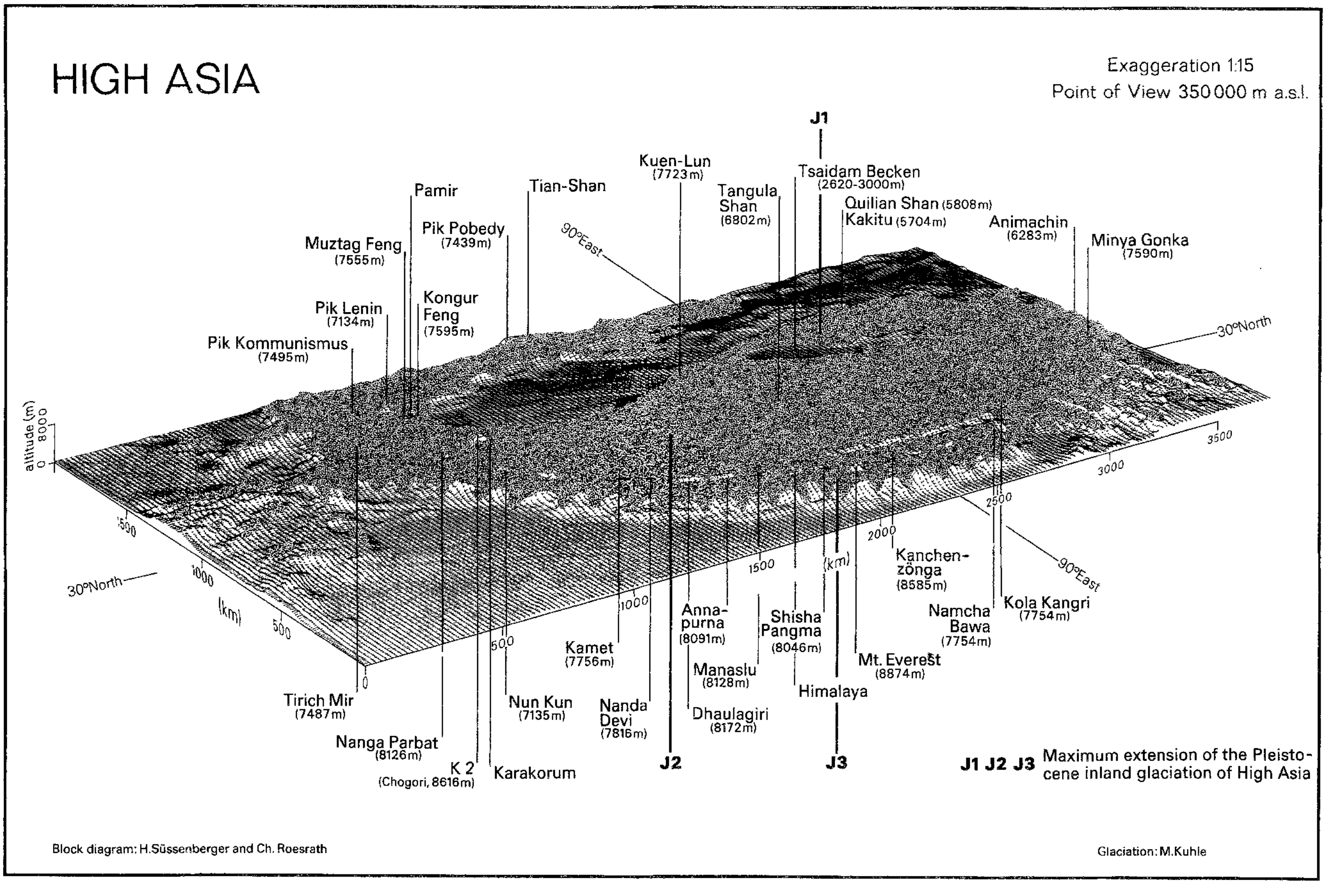

Fig 2 The $2{,}4 \times 10^6$ km^2 continental ice-sheet on the Tibetan highland with its centres I1, I2, I3. Only peaks reaching more than 6000 m to 6500 m project above the glacier surface (exaggeration 15-times)

tion. In the European Alps a decrease in the height of the equilibrium line by 100 m can be produced by a 300 mm/y increase in the precipitation rate or likewise by a 0.6° C decrease in the summer temperature (after Kuhn 1981, 1983). In the subtropical and arid Karakoram area we measured a vertical lapse rate of -0.6 to -0.7° C/100 m (Kuhle 1988a). With approximately the same insolation for glacial times and the present (Berger 1978), a mean depression of the equilibrium line of 1200 m corresponds to a decrease in the summer temperature of 8.4° C, provided precipitation did not change. For a dryer glacial climate the summer temperature depression was probably larger than 10° C.

The Modeling Experiment

Formulation of the Model

The ice sheet model used here (Herterich 1988) allows to calculate the rate of change of ice sheet thicknesses depending on ice flow and on the yearly snow balance on the ice sheet surface. In the so-called shallow-ice approximation (Hutter 1983) the ice flow is determined by the local ice thickness, the slope of the ice sheet surface and the vertical temperature profile within the ice sheet. To calculate the temperature field, the equation for conservation of energy is solved. It includes advection, vertical diffusion and heat production by deformation. At the surface of the ice sheet the (atmospheric) temperature is prescribed and at the ice sheet bottom a geothermal heat flux of $5 \cdot 10^{-2}$ W m^{-2} enters the ice sheet.

In the model, the yearly snow balance is defined by means of three parameters: the height of the equilibrium line h_{SB}, the maximum snow fall b_{max} and that distance Δ h above the equilibrium line, where all the deposited snow is conserved (no melting) and added to the ice sheet. The functional form of the snow balance is plotted in Fig 3. The snow balance is positive and constant ($b = b_{max}$) above heights $z = h_{SB} + \Delta$ h. For heights below $z = h_{SB} + \Delta$ h, the snow balance decreases linearly, with $b < 0$ for $z < h_{SB}$. This highly parameterized form of the snow balance does not contain explicitly the climate variables (temperature, radiation, precipitation) which actually determine the snow balance, but at present no model exists which can provide a more realistic snow balance to be used in ice sheet build-up experiments. However, for the model experiments shown in the next section, this simple parameterized form is probably

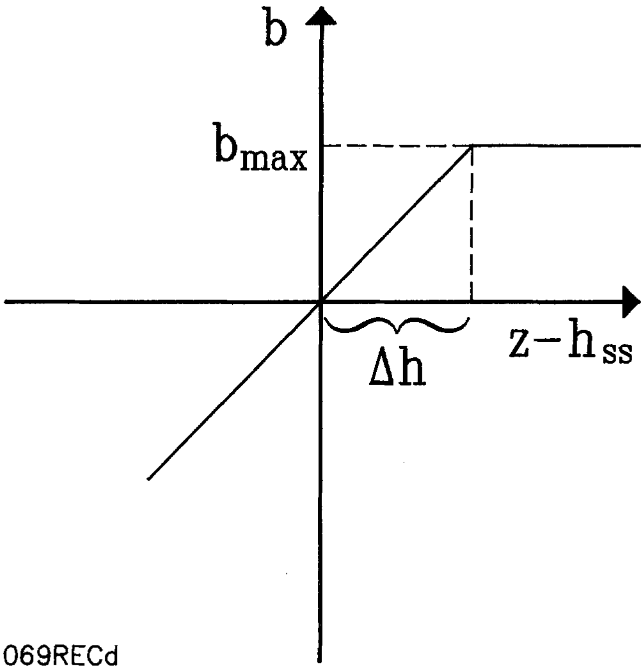

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Fig 3 Graph of the yearly snow balance b as a function of the height above the equilibrium line as used in the model

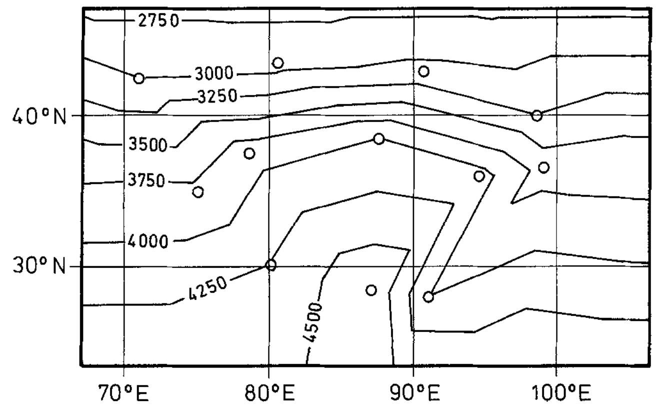


Fig 4 Contour lines (m above sea level) of the glacial "snow surface" (the height of the equilibrium line) in the Tibet area, used in the model. It is fitted to observed heights of the glacial equilibrium line at the positions marked by circles.

sufficient. Model results and comparison with data may in turn show the limits of such a simplified definition of the snow balance.

Modeling Results

The height of the equilibrium line in the model was approximated by a surface (denoted as "snow surface" in the following), composed of planes defined between three points, where the height of the local equilibrium line was available from data. Fig 4 shows the shape of this surface by plotting contour lines of constant height. In wide areas this snow surface lies below the mean height of the Tibetan plateau. Thus in the model an ice sheet starts to build-up in those areas, where the snow balance is positive. The model was then integrated for 10 000 years. During the second half of ice sheet build-up the ice flows also into areas which lay below the snow surface initially. With increasing ice thickness, these areas turn into regimes with a positive snow balance. For this build-up experiment, the maximum snow-fall was $b_{max} = 100$ mm/y, the surface of the ice sheet reached at a height $\Delta h = 1.5$ km above the local height of the equilibrium line. The choice of b_{max} assumes dryer conditions during glacial times, as compared to the present. In S Central Tibet, at Lhasa station in 3740 m height, the yearly mean precipitation is 440 mm/y, with lower values in Gyantse (4000 m, 230 mm/y) and Jomosom (2800 m, 270 mm/y). Both places lie in a dryer area at the southern margin of Tibet. At the W boundary of the Tibetan plateau precipitation ranges from 70 to 200 mm/y within 1500 to 4000 m height, however, in eastern parts of Tibet, precipitation rises up to 1000 mm/y. A value of 400 m/y seems to be a representative precipitation value for the whole plateau. In a dryer glacial climate a value of 200–100 mm/y may not be unrealistic.

In Fig 5 the resulting margin of the model ice sheet is shown after 10 000 model years build-up time, using a value $b_{max} = 100$ mm/y. This modeled ice margin should be compared with the positions of the terminations of outlet glaciers, indicated in Fig 5 by circles. In the mean, model margins match the data points, although larger discrepancies occur locally. The largest error occurs at the S boundary of the Tarim depression (36–37° N/80–82° E), where the model produces a strong outbreak from the Tibetan ice sheet, which is not seen in the geologic record. Also in the E part of the Tsaidam depression the model glaciation is more extended than is documented by observations. Finally, the model predicts a finite ice thickness (> 200 m) at a location in the Tsangpo valley (29–30° N/89–91° E), where the geologic record indicates an ice-free area. Fig 6 and 7 show the increase of the glaciated area and ice volume, respectively, as a function of time. While the ice volume increases almost linearly, the increase of the glaciated area stays at a more or less constant value during the first 4000 years, with a steep rise afterwards. The reason is the onset of ice flow, which is proportional to the fifth power of the ice thickness. The ice sheet can now reach areas, which had a negative snow balance at the time, when the model integration started. With a value of $b_{max} = 100$ mm/y ice thicknesses can at most reach 1 km in areas where the ice flow is negligible. These model thicknesses seem to be much smaller than those thicknesses, which were derived from the geologic record (see above). A possible explanation could be, that the real ice sheet may have had more than 10 000 years time for ice build-up or that precipitation rates were larger than assumed in the model.

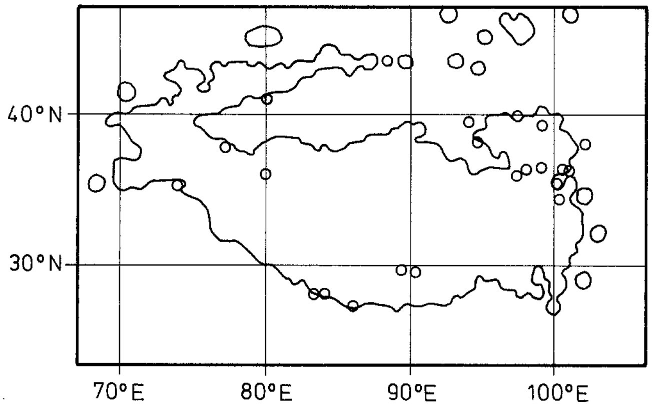


Fig 5 Margin of the model glaciation after 10 000 years ice sheet build-up (solid line) and positions of observed moraines (circles), which mark the maximum extend of the ice age glaciation.

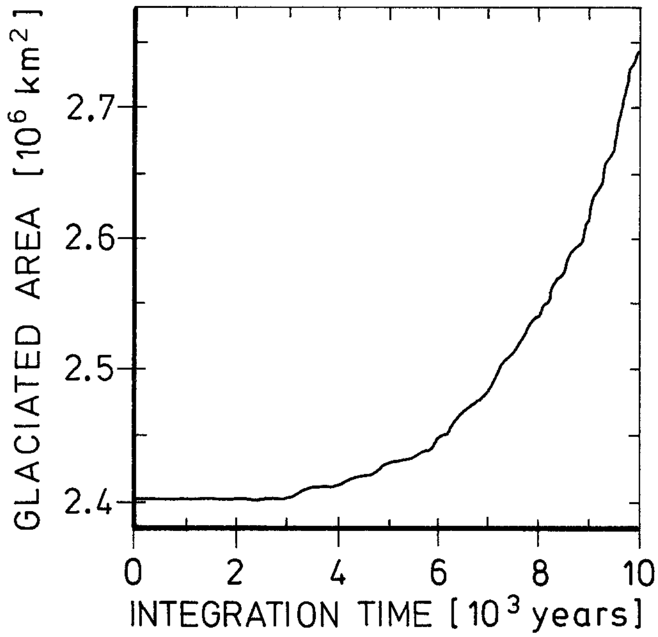


Fig 6 Glaciated area (10^6 km^2) as a function of time (ky), predicted by the model

Discussion

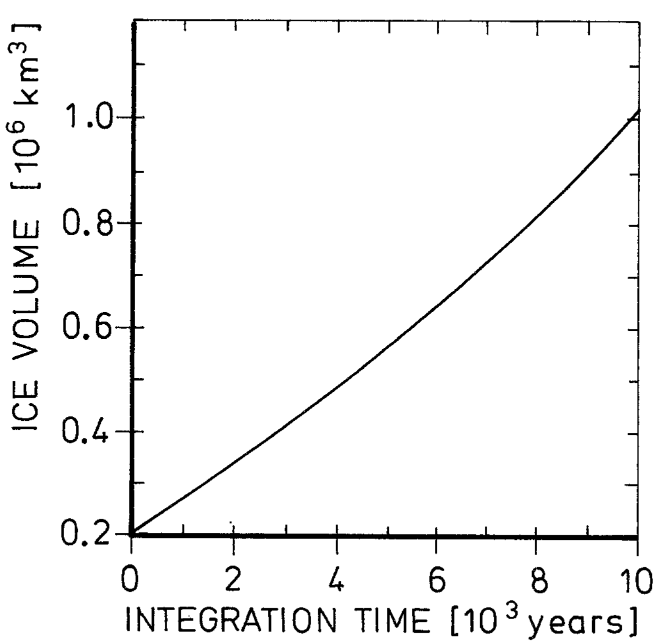
As shown in Fig 5, the available data that can be checked with the model glaciation are rather sporadic. Even these few control points provide hints how the model could be improved in order to fit the data. On the other hand, the model results can be taken as a guide for future expeditions, since the modeled ice sheet margin marks areas, where geologic moraine deposits may be expected.

This close interaction between modeling and measurements is a basic principle in scientific work. Following the principle, the next modeling step will be to replace the constant parameter value b_{max} (maximum snow fall) by values which depend on position. In an inverse modeling approach, b_{max} can be determined such that the model glaciation matches the observed boundaries. The model run with a constant b_{max} (100 mm/y) yielded an outbreak from the Tibetan ice sheet into the Tarim depression, which has not been observed. This outbreak can be inhibited by reducing b_{max} in this area.

Another significant test area is the Tsangpo valley in southern Tibet. The geologic record indicates that this area was ice free. Near Lhasa the glacial outlet glaciers from the Transhimalaya did not reach down to the Tsangpo valley. They terminated in 3900 to 4200 m height, 100 to 400 m above the valley bottom. The model, however, predicts an ice thickness of 200 m at this place, suggesting a lower precipitation than was assumed in the model run. In North Tibet the Kuen Lun and the Quilian Shan were connected by an ice covered area (39–40° N/91–95° E), which is absent in the model. This probably means that the maximum snow fall was larger than 10 cm/y in that area. This rather preliminary comparison between data and modeling results suggests a glacial humidity gradient from NW to E and from S to NE with corresponding values for b_{max} (upper limits for the maximum snow fall) ranging from 50 to 200 mm/y. The inverse determination of b_{max} as a function of position will be the subject of a subsequent paper, including a thorough model sensitivity study with respect to the parameters of the yearly snow balance (h_{sg} Δ h, b_{max}) in the model.

The model results presented above, support the hypothesis for the existence of a Tibetan ice sheet during Pleistocene. Provided the height of the glacial equili-

Fig 7 Model ice volume (10^6 km^3) as a function of time (ky)



brium line has been reconstructed correctly large areas of the Tibetan plateau were above the snow surface and the model then predicts the build-up of an ice sheet within 10 000 years, assuming moderate rates of precipitation. An integration time of 10 000 years is probably a rather short period for ice sheet build-up. Northern Hemisphere ice sheets may have had build-up times of the order of 50 000 to 100 000 years, as is documented in the $\delta\,^{18}O$-record of deep-sea sediments (Shackleton and Opdyke 1973). Thus the build-up time of the Tibetan Ice Sheet was probably also longer than 10 000 years, reaching ice thicknesses in excess of only 1 km, which were predicted by the model. As was mentioned above, the geological evidence indeed yields ice thicknesses larger than 1 km at some places.

The possible existence of a Tibetan ice sheet would have some implications for the interpretation of the $\delta\,^{18}O$-record in terms of changes of the global ice volume. The total amount of additional ice volume of about $50 \cdot 10^6$ km^3 18 000 years BP (Oerlemans and van der Veen 1984) corresponds to a global sea-level change of approximately 130 m. Taking the model ice volume of roughly $2 \cdot 10^6$ km^3 resulting from a 10 000 years build-up, the sea-level would be reduced by 5 m. This is probably a lower limit, since the real ice volume could have been considerably larger. The contribution of the Tibetan ice sheet to the $\delta\,^{18}O$ of ocean waters, however, can be larger than is suggested by simply considering the rather small ratio of the ice volume of the Tibetan ice sheet to the global ice volume. Changes in ice volume of the two major Northern Hemisphere ice sheets (Laurentide and Fennoscandian) were connected also to changes in their lateral extend, which in effect tends to keep the mean $\delta\,^{18}O$ in ice rather constant (around -30‰). Thus, to a first approximation, the deep-sea $\delta\,^{18}O$-record can be taken to be proportional to the changes in the global ice volume. The Tibetan ice sheet, however, will more or less keep the same glaciated area and volume changes are mainly due to variations in ice thickness. This induces a strong change in the mean $\delta\,^{18}O$ of ice during ice sheet build-up (-10‰ km^{-1}), which will introduce a non linearity into the relation between the $\delta\,^{18}O$-record and the global ice volume (Mix and Ruddiman 1984).

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