solubilized by using a neutral detergent and sonication treatment. The Klenow polypeptide in the cell extract was purified to homogeneity in a single step immunoaffinity column chromatography. The purified enzyme showed specific activity identical to that of the commercially available product.

The availability of such a clone would be of immense help in designing strategies for site-directed mutagenesis which is undeniably the most effective approach for identifying the regions in the enzyme molecule that take part in different steps in DNA synthesis.

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RESEARCH COMMUNICATIONS

Enigma of the negative $\delta^{18}O$ pulse at the last glacial maximum in the Arabian Sea

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Sarkar *et al.*[1] had reported a negative excursion of up to 1‰ in the $\delta^{18}O$ of planktonic foraminiferal species at the time of last glacial maximum (LGM) about 18 kyr ago from an east Arabian sea core SK-20-185. They attributed this spike to increased influx of low salinity water from Bay of Bengal through intensified westward coastal circulation caused by stronger NE winter monsoon at LGM. Krishnamurthy[2], however, opined that this negative excursion in $\delta^{18}O$ could be accounted for by the increase in sea surface temperature (SST) resulting from reduced upwelling due to weakening of the summer SW monsoon at the time of LGM. Here we use the limited available data from the east Arabian sea to show that the seasonal abundance pattern of the planktonic species is inconsistent with the mechanism invoked by Krishnamurthy[2]. In view of problems with the mechanisms proposed both by Sarkar *et al.*[1] and Krishnamurthy[2] for the observed $\delta^{18}O$ excursion, we put forward a new mechanism which invokes increased discharge of melt-water from the Tibet and Himalayas into the Bay of Bengal in response to a warming event around the time of LGM.

SARKAR *et al.*[1] have reported variations in the oxygen isotope content of four planktonic foraminiferal species (*G. menardii, G. ruber, O. universa and G. sacculifer*) from an east Arabian sea core SK-20-185 showing a negative $\delta^{18}O$ spike of up to 1‰ at the time of last glacial maximum (LGM) about 18 kyr ago (Figure 1). Alongside of Figure 1 is also shown the measured ^{14}C dates on their core which suggest a time interval of 22–18 kyr BP for the negative $\delta^{18}O$ spike. However, if a uniform sedimentation rate ($2.2\ \mathrm{cm\,kyr^{-1}}$) obtained by considering all the radiocarbon dated levels is assumed, as was done by Sarkar *et al.*[1], the period of $\delta^{18}O$ spike would correspond to around 18 kyr. They[1] cited this spike as an evidence for increased influx of low salinity water from the Bay of Bengal to the core location (10 N, 71°50′ E) through the intensified westward coastal circulation due to a stronger NE winter monsoon at LGM. This interpretation has already provoked a comment by Krishnamurthy[2] who has opined that the negative pulse in the $\delta^{18}O$ of up to 1‰ could be accounted for by the weakening of the summer monsoon and the

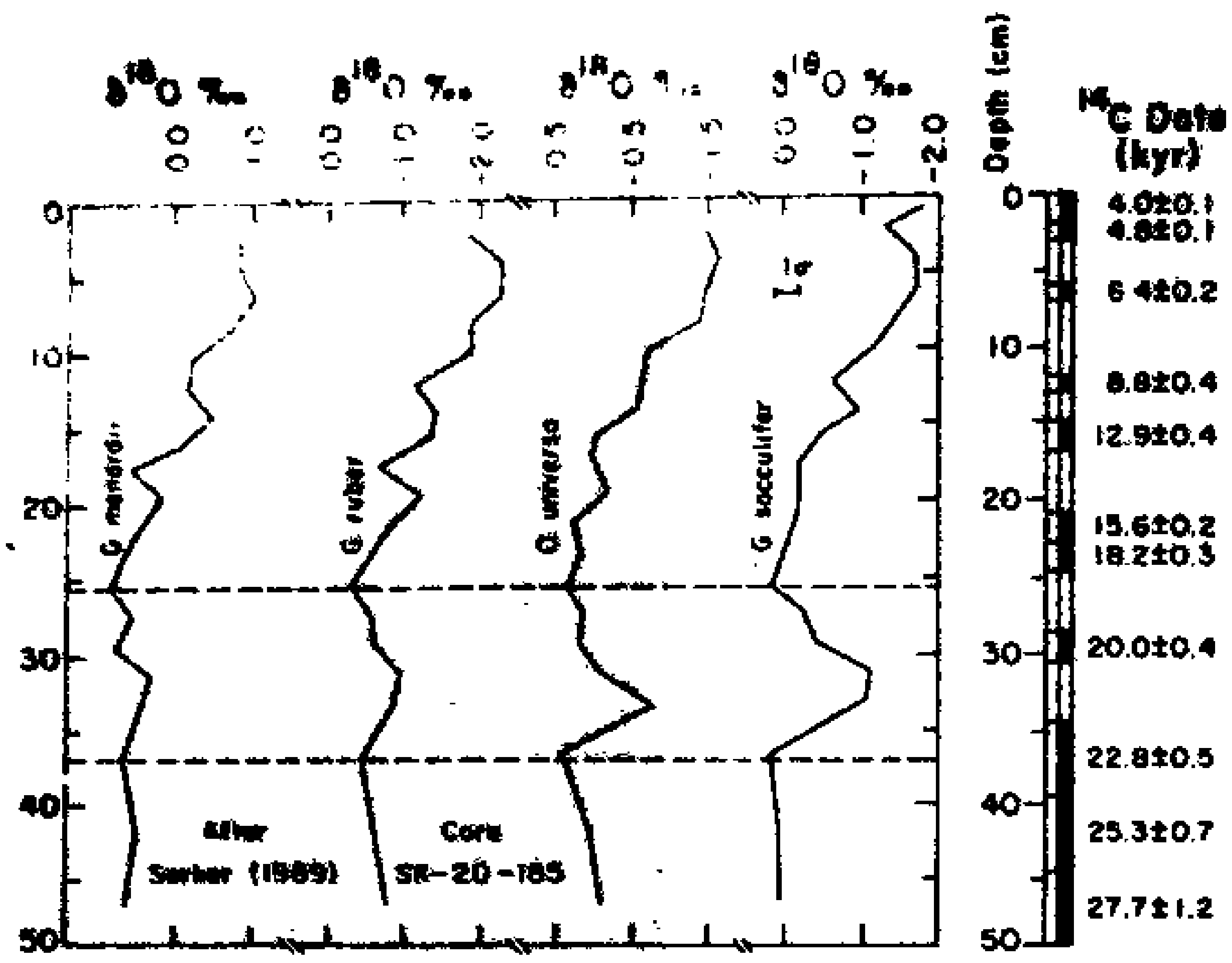

Figure 1. Oxygen isotope data of sediment core SK-20-185 from east Arabian sea, showing a negative spike of up to 1‰, suggestive of freshwater influx, during 22–18 kyr. Radiocarbon dates of core samples are shown alongside. Data from Sarkar[18].

resulting increase in SST due to reduced upwelling in the Arabian sea at the time of LGM. This argument was put forward on the basis of the difference in $\delta^{18}O$ between two planktonic foraminiferal species (*G. sacculifer* and *G. ruber*) analysed by Sarkar *et al.*[1] and difference in the seasonal distribution of their peak abundance in the present day western Arabian sea[3].

The present day SSTs in the Arabian sea steadily increase from February, reach a maximum in April–May, start decreasing from June and reach another maximum in October[4]. According to Krishnamurthy[2], due to the weakened SW monsoon at LGM the region could have behaved like middle and high latitudes where SSTs increase steadily through the summer, so that both *G. ruber* and *G. sacculifer* having their abundance peaks in summer and late summer/early autumn respectively[3] would have secreted their shells at 2°–4°C higher temperature, and hence accounted for their negative spike of 0.5–1.0‰ (for *G. ruber*) and 1.0‰ (for *G. sacculifer*). However, in contrast to the speculation of Krishnamurthy[2] on the seasonal abundance pattern of the two species, limited data available[5] from the eastern Arabian sea suggest that *G. ruber* shows highest abundance during pre-monsoon (May) and *G. sacculifer* in February–March.

Further, if the mechanism of Krishnamurthy[2] were operative, one should see a negative $\delta^{18}O$ spike of 1‰ or more for *G. sacculifer* around LGM, in other high resolution analyses of sediment cores from the Arabian sea especially from the active monsoonal upwelling areas off the coast of Somalia and Oman. The core CD-17-30 (19°56′N, 61°39′E) off the coast of Oman, having a sedimentation rate of 7.7 cm kyr^{-1}, shows only a 0.5‰ negative $\delta^{18}O$ excursion[6] for *G. sacculifer* at LGM, a value consistent with a 2°C increase of SST associated with reduced upwelling and reduced strength of summer monsoon[7] at LGM. Since this high sedimentation rate core is derived from a known active monsoon upwelling region in the Arabian sea in the vicinity of the Gulf of Aden where the *G. sacculifer* abundance is reported[3] to peak in late summer/early autumn, the 0.5‰ negative excursion observed in this core for *G. sacculifer* at LGM should be considered an upper limit ascribable to the mechanism proposed by Krishnamurthy[2]. Thus, in the absence of (i) observational data on the seasonal distribution of the foraminifera *G. sacculifer* and *G. ruber* from the east Arabian sea in support of the speculation of Krishnamurthy[2] and (ii) absence of evidence for a comparable magnitude of the negative $\delta^{18}O$ excursion from other parts, particularly from active monsoonal upwelling region of the Arabian sea, we suspect that the $\delta^{18}O$ spike observed in SK-20-185 is not wholly ascribable to the mechanism proposed by Krishnamurthy[2].

In view of problems with the mechanisms so far proposed by Sarkar *et al.*[1] and Krishnamurthy[2] to explain the observed $\delta^{18}O$ spike at LGM in SK-20-185, we now propose another mechanism which invokes discharge of glacial melt-water from the Tibet and Himalayas into the Bay of Bengal. This melt-water would reach the core location in the east Arabian sea by westward advection from the Bay of Bengal through coastal circulation as proposed by Sarkar *et al.*[1].

Possibility of increased discharge of the isotopically light Himalayan melt-water is indicated by the observation of a warming event during 22–18 kyr inferred from the Dunde ice core (37°6′N, 96°24′E; elevation 5326 m) record of isotopic and chemical measurements (Figures 2*a–e*) of Thompson *et al.*[8]. It is noted that during the time interval 22–18 kyr, $\delta^{18}O$ (a temperature proxy) becomes less negative indicating warming, total particles decrease and soluble salts increase. The chronological estimates of Dunde ice cores were based on counting of annual $\delta^{18}O$ and dust peaks up to 117 m depth, where the ice was deposited 4550 years ago. The age of the prominent stratigraphic transition at 129.2 m was tagged to 10,750 years to correspond with age of last glacial Holocene transition in the Camp Century core from northern Greenland[9]. The age of the deeper ice was obtained in analogy with the deep profile from the Dome C ice divide[10] by fitting an empirical power law relationship between the ice layer thickness at a given depth, the accumulation rate and the depth of the ice layer[8]. We also note from Thompson *et al.*[8] that their sampling interval was 3 cm. Therefore, according to their scheme of dating, each 1000 year average in Figure 2*a–e* was apparently obtained from a minimum of 2–3 samples even at the oldest part of the profile (39–40 kyr), progressively increasing to 7–8 samples in 10–12 kyr period. A

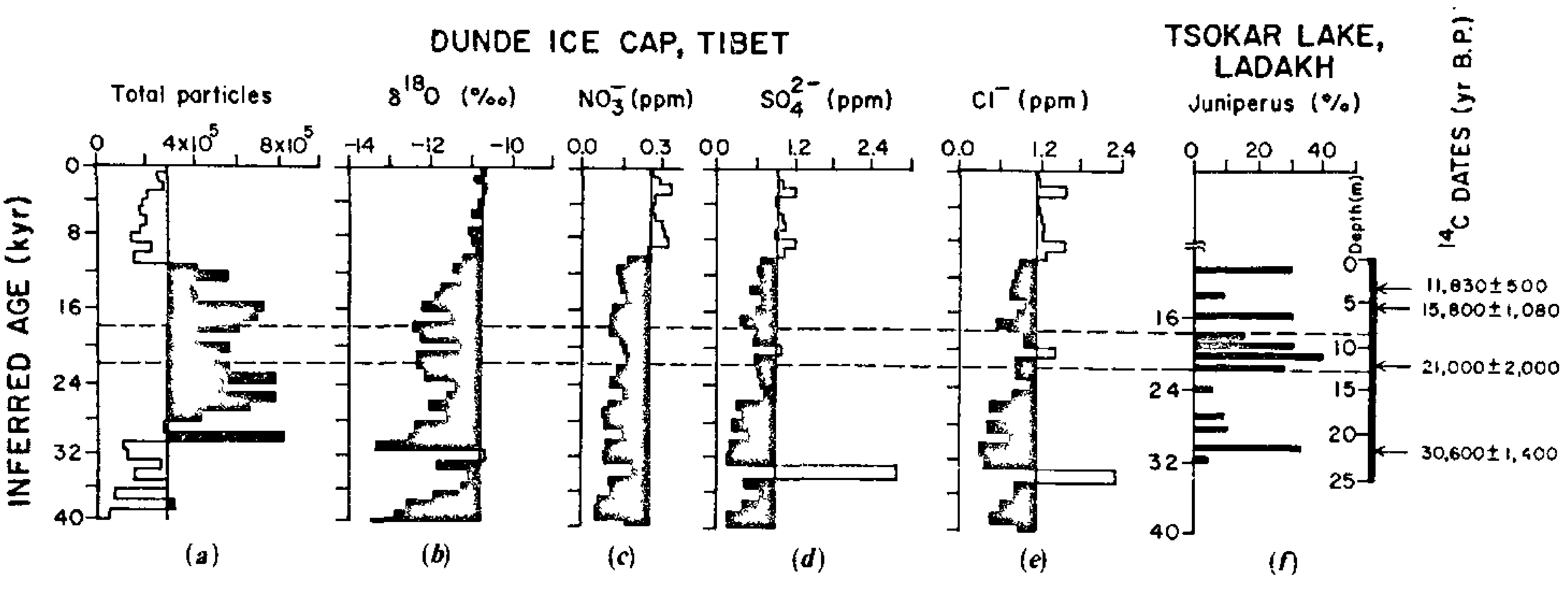

Figure 2. Discrete 1000 yr average of (*a*) dust concentration; (*b*) $\delta^{18}O$; (*c*) NO_3^-; (*d*) SO_4^{2-} and (*e*) Cl^-, in the Dunde ice core (redrawn from figure 4 of ref. 8). Figure 2 *f* gives the *Juniperus* abundance profile from Tsokar lake, Ladakh (redrawn from figure 3 of ref. 11).

similar warming event is indicated from the palaeovegetational data of Tsokar lake (33° 21′ N, 78° E; elevation 4572 m), Ladakh of Bhattacharyya[11]. This study reveals (Figure 2*f*) continuation of alpine steppe with periods of expansion of *Juniperus-Poaceae-Uticaceae* communities during 22–18 kyr BP. The chronology of the pollen profile is based on four radiocarbon dates as indicated in Figure 2*f*. It is seen that independent proxy-palaeoclimatic indicators from the two flanks of the Tibetan plateau show, within limits of the dating accuracy of the two profiles, concurrent, warming episodes at 22–18 kyr (shown bracketed by the two dashed lines in Figure 2) and also at 32–28, 16–15 and at 10 kyr.

Evidence for a regional warming event during 22–18 kyr is also provided by the dating of palaeosol horizons from Kashmir by Kusumgar *et al.*[12] from the loess sections at Puthka (34° 14′ N, 74° 21′ E), Burzahom (34 52′ N, 74°47′ E), Dilpur (33°56′ N, 74°57′ E) and Tilsur (33° 52′ N, 74° 57′ E) in the range 19.5 ± 1.9 kyr (Figure 3). Krishnamurthy *et al.*[13] found that the palaeosol at Burzahom had an organic fraction with $\delta^{13}C = -25‰$ which is indicative of the dominance of C_3 type of vegetation that thrives under climatically optimal conditions. This is in contrast to the $\delta^{13}C = -16‰$ in the loess layer below, indicative of C_4 type of vegetation characteristic of arid climate. The inference is that soil formation during 22-18 kyr BP occurred under the influence of warm and humid climate as opposed to cold and arid climate when loess deposition was enhanced.

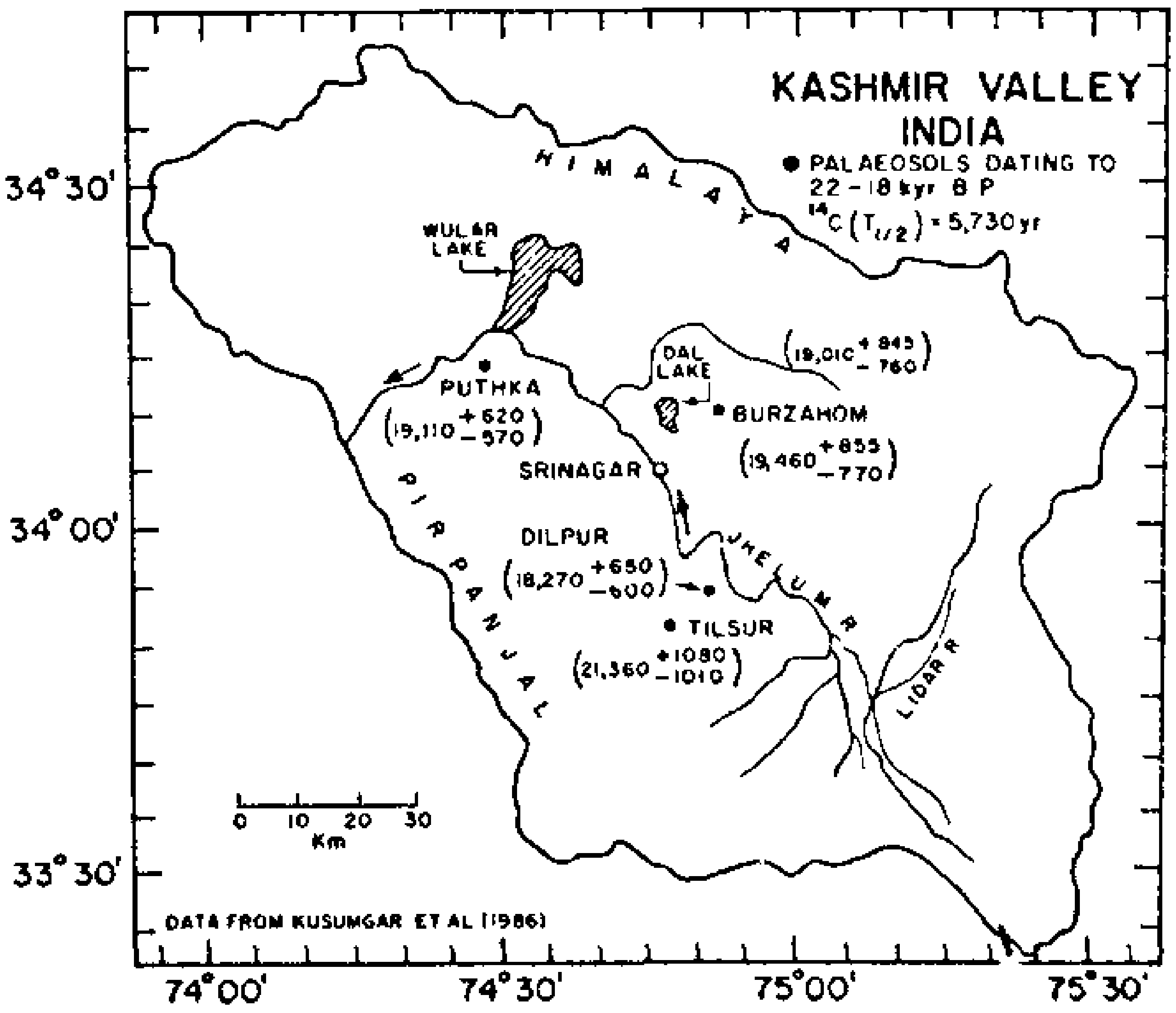

Figure 3. Schematic map of Kashmir valley, India, showing the locations of the palaeosols dating to 22 18 kyr. The radiocarbon dates given by Kusumgar *et al.*[12] based on ^{14}C half-life as 5568 yr are converted to the presently accepted half-life of 5730 yr.

Palynological analysis[14] of a peat bog at Butapathri (36°06′ N, 74° 43′ E) in Kashmir revealed a distinct climatic amelioration at 18 kyr, marked by partial dominance of broad leaved thermophilous elements such as *Ulmus*, *Alnus* and *Juglans*.

We, thus see that palaeoclimatic data from a number of proxy-indicators converge to suggest a warm (warm humid?) episode in an otherwise glacial climate in the time interval 22 18 kyr in the broad region extending from Kashmir to northwest Tibet.

A question that naturally arises concerns the reliability and contemporaneity of the evidence from different locations. We have seen that the chronology of the Dunde ice cap data is based on an empirical power law fit to the age data obtained by counting of annual layers and by tagging the prominent stratigraphic transition to 10,750 yrs. In such a scheme, the age error on deeper (and hence older) parts of the profiles increases progressively. Therefore, an age error of 2–4 kyr at 22–18 kyr interval is not unlikely. Similarly, different radiocarbon ages on different fractions of soil organic matter from the same sample have also been reported[15,16]. We, however, believe that the contemporaneity of inferred warming at 22–18 kyr from a large number of different locations spread over a wide geographical area in the Indo-Tibetan region may not be a fortuitous coincidence, but may represent a definite event at this time. At this lower end, the time bracket of 22–18 kyr also encompasses the period of LGM and it may, therefore, be tempting to correlate it with the LGM. Until more reliable independent dating evidence becomes available, the contemporaneity of events as indicated above may be accepted but with the possible sources of error kept in mind. In the above discussion we have ignored the problem associated with divergence of radiocarbon and calendar years, because most of palaeoclimatic data in this note as also in the literature, including the assignment of LGM to 18 kyr, are based on radiocarbon years.

Gupta *et al.*[17] have modelled the mixed layer of the Bay of Bengal as a well-mixed box and have estimated the degree of change required in the quantum of river discharge and/or its isotopic composition to cause the observed negative $\delta^{18}O$ spike in core SK-20-185. The model calculations show that if the increased run off from the east flowing peninsular Indian rivers had been responsible for the observed spike, the rainfall during NE winter monsoon at the time of the LGM should have increased by a factor of 5–10 depending on its $\delta^{18}O$ value. On the other hand, a melting of $< 10\,\mathrm{cm\,yr^{-1}}$ of the accumulated snow during the 22–18 kyr would easily cause the lowering of $\delta^{18}O$ of the mixed layer of the Bay of Bengal and account for the observed magnitude of the spike.

We have no direct observational evidence yet for the melt-water spike in the Bay of Bengal. There is, however, strong observational evidence for a warming event during the 22–18 kyr period through several independent proxy climate indicators from widely separated areas in the Indo-Tibetan region as discussed above[8,11,12,14,17]. This warming event could have provided the melt-water needed to explain the observed spike. Evidence for melt-water of the Indo-Tibetan origin can be obtained by a high resolution investigation of sediment cores from suitably selected locations in the Bay of Bengal which would have been directly influenced by such an influx of melt-water at LGM. A correct identification of the source of the spike is important for palaeoclimatic reconstruction of this region and may have profound implications for future modelling studies.

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Metal levels in zooplankton from Hooghly estuary, India

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We have measured the levels of Fe, Cd, Ni, Pb and Co in zooplankton collected from Chemagari creek of Sagar Island of Hooghly estuary. Among the metals studied Fe showed maximum concentration. The concentration of Pb and Co was below the level of detection limit throughout the season. The concentration of Fe, Cd and Ni in zooplankton varied from 1470 to 12051 ppm, 6 to 96 ppm and 4 to 29 ppm respectively. Negative