rates of bacterial metabolism may be limited by the diffusion of electron acceptors through a large thickness of sediment[21]. Thus, although the absence of benthic activity[22] and low metabolic rates in the water column[23] may diminish the decay of metabolizable nitrogenous material, it seems certain that the upper unit contains organic matter of Recent (or at the very oldest, late Pleistocene) origin. Radiocarbon ages support a Recent origin[4]. Recycled spores and pollen have been identified in RISP sediments[2,11], but the C/N ratios of plant material (13-26) are much larger than the C/N values in the upper unit, and recycled terrestrial material cannot be the main organic matter source.

The diagenetic gradients indicate that deposition of the upper unit has occurred postglacially, with fresh organic matter being incorporated during the reworking of older sediments. Thus the geochemical data are consistent with the occurrence of a reworking event which mixed Pliocene and Pleistocene fauna with the Miocene forms[4], but not with the existence of an *in situ* Miocene succession[1,2]. No reworking of the upper unit has occurred following deposition at the present site, otherwise no depositional or diagenetic variations would occur. In this respect, we agree with the conclusions based on ^{10}Be measurements, although these imply a pre-Quaternary origin at the latest for the upper unit[24]. Such a prolonged period of non-deposition is incompatible with the existence of reduced S species, and nitrogenous organic matter, at the sediment-water interface. Comparatively large ^{10}Be ages may result from diminished rates of ^{10}Be deposition at high latitudes and in the presence of an ice cover[25].

As the two units are so similar in detrital mineralogy, their main chemical differences arise as a result of alteration by *in situ* diagenesis which differs only in degree and not in direction. The extent of alteration by diagenetic sulphate reduction is mainly controlled by metabolizable organic matter concentrations and a fundamental difference between the two units is the lower organic C content of the upper unit, even though the diagenetic loss of organic C is not yet completed here. Hence, the lighter colour of the upper unit is probably due to a smaller content of black, iron monosulphides arising from both a lower initial organic C content and unfinished sulphate reduction. Complete diagenetic alteration of the upper unit would result in a sediment with less organic C and S than the lower unit, although the progressive C/N and C/S trends would suggest that the two units would ultimately be very similar in these respects.

The total S, C/S and C/N trends appear to pass through the iron oxide layer (at 18-cm depth) without major perturbation, but by 25 cm have closely approached the values at greater depths. The depth at which diagenesis is mostly complete thus lies only a little below the iron oxide layer and hence there are no unequivocal signals of depth-age relationships in the lower unit. However, the small but irregular trends of increasing total S and C/N values would be consistent with continuity of diagenesis through the iron oxide layer, and hence with a post-late Pleistocene origin for the whole core.

The iron-stained layer between the two units is a major chemical discontinuity whose temporal significance is unclear. No samples at this depth were analysed, but the small kink in the C/S ratio at 15-16 cm suggests a possible loss of S which, together with the presence of iron oxides, implies a discontinuity in sedimentation which allowed the oxidation of sulphidic sediments by exposure to oxygenated seawater. The oxidation of sulphides may proceed quite rapidly under these circumstances and numerical models for the oxidation of sulphidic sediments by the diffusive penetration of dissolved oxygen into the RISP sediments indicate that a 1-cm oxide layer could be produced in as little as 100 yr.

The diagenetic chemistry clearly demonstrates that the RISP sequence is not wholly Miocene and that the upper unit is probably of Recent origin. Unfortunately, there are no clear indications as to the age of the lower unit. Our observations can be reconciled with the presence of Miocene, also Pliocene and Pleistocene, fauna if the upper unit was formed by reworked material (similar to that in the lower unit) which was redeposited following suspension in the water column (which would allow an autochthonous organic matter contribution). There are no chemical features which are inconsistent with a post late-Pleistocene origin for the whole core.

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Late-Quaternary climatic change on the American North Pacific Coast

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Temporal changes in solar radiation caused by variations in the Earth's orbit figure prominently in current Quaternary climate theory[1-4]. Experimental design of atmospheric circulation models emphasizes the importance of solar radiation in the heating of interglacial land and ocean surfaces at different seasons of the year, whereby monsoon-type circulation—seasonal shifting of regional cyclonic and anticyclonic centres—develops, while during glacial ages, seasonality is greatly modified by the influence of ice sheets[5-7]. We investigated the late-Quaternary climate of the North Pacific, where according to modelling, solar radiation in the early Holocene at the time of the summer solstice is high and in the late Holocene is relatively low. We compared quantitative temperature and precipitation estimates from southern Alaska, obtained by application of transfer functions to fossil pollen records, with estimates from western Washington[8] and southern British Columbia[9]. Data extending over >10,000 years show a broadly consistent pattern of climatic change in general agreement with predicted variations in solar radiation and their effect on atmospheric circulation and seasonal duration of pressure systems over the North Pacific Ocean. In the early Holocene, as monsoon-type circulation became established with melting of glaciers, the subtropical North Pacific anticyclone annually regulated climate for a longer period at higher latitudes than at present, so that warmth and dryness (high summer radiation) increased in southern Alaska. The Aleutian low-pressure centre, locus of cyclonic storm activity, intensified during the late Holocene, resulting in colder and more humid coastal climate (low summer radiation) over much of the year and increased frequency of glacier growth in the cordillera.

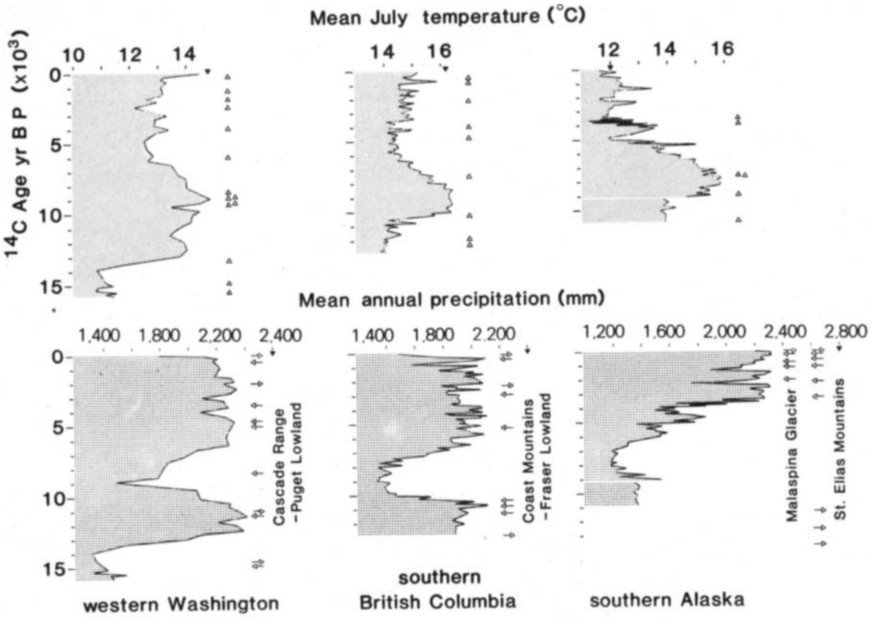

Fig. 1 Late-Quaternary mean July temperature and annual precipitation estimated by use of transfer functions for sectors of North America located in western Washington (47° 49′ N, 124° 12′ W)[8], southern British Columbia (49° 15′ N, 122° 20′ W)[9], and southern Alaska (60° 01′ N, 141° 57′ W). Data are drawn to a uniform time scale with chronology indicated by radiocarbon ages (Δ) and assuming constant sedimentation rates between ages. Markers on temperature and precipitation scales are modern means. Arrows pointing left next to precipitation curves signify times of glacier advance, whereas retreat is indicated by arrows pointing right; vertical lines between arrows indicate extended intervals of glacier activity.

Under the influence of a trough of low pressure located in the Aleutian/Gulf of Alaska region, climate today along the American north-west coast is mostly stormy, cold and rainy[10]. In summer, as the subtropical anticyclone develops over the North Pacific Ocean, the Aleutian low is weakened and displaced northwestward by the build up of high pressure. Some weeks of relatively stable conditions ensue in the western conterminous United States and southwestern Canada, providing a contrast with the weather during the colder part of the year. Although summer months in southern Alaska and northern coastal British Columbia are driest, influenced by the seasonal pressure change, cyclonic storms, accompanied by heavy precipitation and cloudiness, continue to strike this sector of the coast.

Seasonal duration and extent of pressure systems over the North Pacific were variable in the late Quaternary, as implied by quantified mean July temperature and annual precipitation estimates for three geographic areas (Fig. 1). Data from southern Alaska serve for comparison with data from the Olympic Peninsula of western Washington[8] and the Coast Mountains of southern British Columbia[9]. All three data sets were generated by two independent regression equations[8] that relate modern composition of pollen in surface samples from the Pacific coastal forest and tundra regions to recorded temperature and precipitation values at sites between the Aleutian Islands and California. Applied to radiocarbon-dated fossil pollen records in the three areas, the equations yield estimates of late-Quaternary climatic parameters. Transfer functions used were applied within the limits of the calibration data set.

Southern Alaska data (Fig. 1) are derived from fossil pollen contained in two sections of a muskeg located near Munday Creek on the Gulf of Alaska, ~15 km north-west of Icy Cape in the Malaspina Glacier district. One section extends to ~9,000 yr BP[11] whereas the other reaches beyond to 10,820 yr BP[12]. Mean July temperature is near 14 °C at the end of the Pleistocene (10,000 yr BP) and in the early Holocene, reaching a maximum close to 16 °C at ~8,000 yr BP and then decreasing to 12 °C at ~5,000 yr BP, with fluctuations around this value until the present. Annual precipitation, just under 1,400 mm at the beginning of the record, reached a minimum near 1,200 mm, coinciding with the temperature maximum at ~8,000 yr BP, and thereafter increased to almost twice as much during the past four millennia, while temperature fell to its lowest level of the entire time span.

Temperature and precipitation trends expressed by the curves represent a pattern, repeated to the south in British Columbia and Washington (Fig. 1). All sites show early Holocene temperature maxima and precipitation minima, whereas in the late Holocene (after 5,000 yr BP), as in the late Pleistocene, higher precipitation is coincident with lower temperature. Increased warmth and dryness during the early Holocene suggest extended dominance by the subtropical Pacific anticyclone; conversely, before and after this interval, when precipitation amounts are higher and temperature values lower, polar Pacific cyclones became effective in generating storms that increased in frequency over a greater part of the year. Changes associated with the centres of atmospheric pressure were influenced to a great extent by land-ocean temperature contrasts, latitudinal temperature gradients and topography. Changes follow closely predictive general circulation models in which monsoon-type climate is related to variations in seasonal solar radiation intensity[5–7].

Storm activity, implied by the fossil evidence, was most intense in the late Holocene in southern Alaska, where precipitation increased by some 1,200 mm and temperature fell ~4 °C. Particularly wet climate effecting generation of the muskeg studied near Icy Cape is indicated by an unusually high deposition rate (42 mm yr^{-1}, assuming no compaction) for 150 cm of unhumified peat formed between 3,860 and 3,500 yr ago; before this, under a warmer, drier climate between 9,000 and 7,700 yr ago, rate of deposition of 25 cm of highly humified peat was much lower (2 mm yr^{-1}). Rapid peat deposition close to 3,500 yr ago has been found elsewhere in southern Alaska[13].

This pronounced climate change occurred at the time of the Neoglaciation, during which colder wetter conditions, giving rise to excessive snowfall in the lofty St Elias Mountains, caused glacier regeneration (Fig. 1) between 3,300 and 2,100 yr ago, 1,250 and 1,030 yr ago and in recent centuries, after an extended interval of early Holocene recession[14–18]. During the Neoglaciation, ice flow from the St Elias Mountains contributed to the formation of the huge piedmont lobe of Malaspina Glacier, which spread across the coastal lowland and continental shelf to the Pacific. Episodes of ice advance in the Malaspina Glacier district occurred over the past two millennia, with maxima dated between 1,600–1,200 and 830–560 yr ago and in the past few centuries[19–23].

Malaspina Glacier, covering some 2,200 km^2, rests mostly below sea level[24] along a part of the Gulf of Alaska coast that, under the warmer, drier climate of the early Holocene, was probably an embayment. Located between the eastern end of the Robinson Mountains at Icy Bay and the Deception Hills at the mouth of the Alsek River, the embayment extended along >200 km of coastline, including present-day Icy and Yakutat Bays, before the Neoglaciation. Higher temperature and lower precipitation estimates make it unlikely that mass balance of Malaspina Glacier in its present lobate form could have been maintained in the early Holocene.

Greater frequency of storms passing across southern British Columbia and Washington over the past six or seven millennia is reflected by the estimates of increased precipitation in this part of the coast (Fig. 1). The increase, while temperature steadily decreased, inaugurated late Holocene glacial episodes at ~5,000 yr ago[25-31]. During the early Holocene, except for isolated instances of minor glacial advance in the Cascade Range[32-34], glaciers were inactive. Climate, under control of the subtropical anticyclone, was apparently too warm and dry for glacier growth. In the late Pleistocene (15,000-10,000 yr ago), a major advance occurred in the Puget Lowland in a cold, dry climate ~14,500 yr ago[35], followed under wetter but less cold conditions by minor advances in the Fraser Lowland and North Cascades close to 11,000 yr ago[36-38] and in the Coast Mountains ~10,500 yr ago[39]. Late Pleistocene glaciation of the Puget Lowland has been attributed to northward movement of storm tracks, causing heavy snowfall in the glacier source region in the Coast Mountains of British Columbia, coincident with decay of the Laurentide ice sheet[40].

In the long-term, late-Quaternary climate fluctuations along the American north-west coast, keyed to shifting dominance of atmospheric pressure cells over the North Pacific, are outlined by the pattern of temperature and precipitation estimates. They conform with temporal changes in seasonal radiation intensity predicted by experimental modelling[5-7], previously brought out by fossil pollen studies in north-west Canada[41]. In the absence of tighter radiocarbon-dated time control over extended intervals in the records, assessment of short-term fluctuations of the estimates (of the order of centuries) is not justified. Rates of response to climate change, as well as speed of migration northward following deglaciation of the British Columbia and Alaskan coasts, by plant species used in the equations, for example, are variables that must enter into the timing of events. These may be significant, but further dating and replication are required to define them.

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Evidence for an early Plio-Pleistocene rainforest expansion in East Africa

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The Plio-Pleistocene palaeoenvironmental history of the Turkana Basin and neighbouring areas in north Kenya and southern Ethiopia has been a matter of considerable interest due to the important palaeoanthropological, palaeoarchaeological and palaeontological discoveries that have been made in this region[1-3]. The consensus is that although environmental oscillations have certainly occurred over the past 4 Myr, Plio-Pleistocene environments were generally comparable to the modern environment of the area[4]. However, the recent recovery of two taxa presently characteristic of the central African rainforest—the anacardiacean tree *Antrocaryon* and the prosobranch gastropod *Potadoma*—from a restricted chronostratigraphic horizon dating at ~3.3–3.4 Myr, at widely separated sites in the Turkana region, suggests that the general pattern of environmental stability was punctuated at this time by a brief but significant rainforest extension in East Africa. This brief humid interval apparently coincides with certain important molluscan immigrations, and can be broadly correlated with the first major episode of climatic deterioration in the Northern Hemisphere.

Recent reviews of Plio-Pleistocene palaeoenvironments in the Omo Valley and Koobi Fora areas of the Turkana Basin, north Kenya, have drawn on sedimentological, palaeontological and palaeobotanical evidence (see, for example, refs 1-4). Although a relatively xeric interval between 2.5 and 2.0 Myr has been postulated[5], the evidence for significant environmental change at this time has been questioned[4]. The general consensus seems to be that throughout the past 4 Myr the Turkana region has been characterized by a suite of terrestrial environments similar to those found at present—dry tropical scrub and grasslands of the 'Sudan savanna' replaced locally by limited gallery forest in the neighbourhood of perennial drainages[4]. However, several recent lines of evidence suggest that the general pattern of overall environmental stability in the latest Cenozoic of East Africa may have been briefly but significantly perturbed by an important humid episode at 3.3-3.4 Myr, an episode which entailed a major expansion of the central African rainforest. The suggestion that a humid episode occurred at this time was originally prompted by the discovery of fruits of the anacardiacean rainforest genus *Antrocaryon*[6] in the 'Brown Sands' locality of the Usno Formation, a short stratigraphic distance above tuff U10, and from Member A1.1 of the Shungura Formation in the lower Omo Valley (see Fig. 1), immediately underlain by tuff A and overlain by tuff B. Tuff U10 is considered to be a lateral correlate of tuff B of the Shungura Formation (F. H. Brown, personal communication), and tuff B itself is now considered a lateral correlate of the Tulu Bor Tuff from the Koobi Fora area[7,8]. The age of this latter horizon is considered to be ~3.35 Myr, on the basis of its stratigraphic position between the immediately overlying Toroto Tuff (firmly dated at 3.32 ∓ 0.02 Myr) and the underlying Lokochot Tuff[9]. On the basis of geochemical comparisons, the Lokochot has been correlated with tuff A of the Shungura, which is known to lie close to the boundary between