The Karakoram Anomaly? Glacier Expansion and the 'Elevation Effect,' Karakoram Himalaya

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In the late 1990s widespread evidence of glacier expansion was found in the central Karakoram, in contrast to a worldwide decline of mountain glaciers. The expansions were almost exclusively in glacier basins from the highest parts of the range and developed quickly after decades of decline. Exceptional numbers of glacier surges were also reported. Unfortunately, there has been no on-going measurement of climatic or glaciological variables at these elevations. The present article examines possible explanations for this seemingly anomalous behavior, using evidence from short-term monitoring programs, low-altitude weather stations, and the distinctive environmental characteristics of the region. The latter involve interactions between regional air mass climatology, its seasonality, topoclimate or 'verticality' effects on glaciers with extreme altitudinal range, climatic sensitivities of heavy versus thin supraglacial debris, and complex temperature distributions in ice masses with ice falls throughout critical elevations. Valley climate stations indicate increases in precipitation over the past 50 years and small declines in summer temperatures, which may indicate positive trends in glacier mass balance. However, the suddenness of the expansions is problematic, as is their confinement to glaciers from the highest watersheds while others continue to retreat. Thermal shifts in ice masses with extreme altitude ranges may be even more critical, leading to an accelerated redistribution of ice mass by elevation.

Introduction

In recent decades most of the world's mountain glaciers have undergone negative mass balance and terminal recessions (Haeberli et al 1999; Oerlemans 2002). Rapid declines are reported throughout the Greater Himalaya and most of mainland Asia (Ageta 2001). These trends have been attributed to global atmospheric warming. From the 1920s to the early 1990s, most glaciers of the Karakoram Himalaya were also observed to diminish, except for some short-term advances in the 1970s and surges. However, in the late 1990s, many central Karakoram glaciers began expanding (Figure 1).

Perennial snow and ice covers about 18,500 km^2 of the Greater Karakoram–Hindu Raj Ranges, and is dominated by large glaciers (von Wissmann 1959). The expansions apply only to the highest watersheds of the central Karakoram. Most glaciers of interest originate above 7000 m and have elevation ranges of over 4500 m. Some descend as low as 2300 m, much lower than in the Greater Himalaya of India and Nepal. Generally, the glaciers differ in size, elevation, and latitude from most glaciers used to track contemporary global changes (Oerlemans 2002; Haeberli 2004; WGMS 2005). Other distinctive features include the regional climate, glacier nourishment, and extreme vertical gradients. Heavy debris covers, steepness, and thermal characteristics distinguish the ice masses themselves. However, there is another unfortunate 'anomaly' for understanding these glaciers and their changes: the absence of on-going measurement programs for climatic and glaciological variables, necessary to identify changes in mass balance and other properties influencing ice behavior. In part this reflects the high costs and difficulties of working in the region, in part the need to develop alternative methods from those most widely deployed elsewhere (Young and Hewitt 1993). All of this contributes to the comparative neglect of Karakoram glaciers in the recent literature. The present article first describes the observed changes, and then provides an overview of what is known of conditions in the Karakoram that may help to explain the apparent anomaly and to identify future monitoring needs.

Recent glacier changes

In surveys between 1997 and 2002, 13 glaciers of intermediate size—10 to 20 km in length—were found to be advancing, as were 16 disconnected tributaries of large glaciers (Table 1). Some of the larger glaciers—40 to 70 km in length—exhibited 5 to 15 m of thickening over substantial ablation zone areas, locally more than 20 m (Figure 2). Along 8 glaciers, renewed lateral moraine building involved thicker, more active ice overriding moraines that were ice-free for 50 years or more (Figure 3). Exceptional numbers of glacier surges were reported, 6 of a possible 9 confirmed between 1986 and 1996—the most in any decade since the 1830s (Hewitt 1998b). Thickening appears irregular in amount and incidence, with wave-like zones of higher ice moving down the glacier. The expansions developed quickly where thinning and recession had been observed between 1985 and 1995 (Hewitt 1998a).

The changes detailed in Figure 1 and Table 1 are based on ground observations between 1985 and 2001. They were first identified at Bualtar, Barpu, Hispar, and

FIGURE 1 The central Karakoram Himalaya. The map summarizes changes observed in the period 1997–2002 (see also Table 1), and surges since 1986. (Map by Pam Schaus)
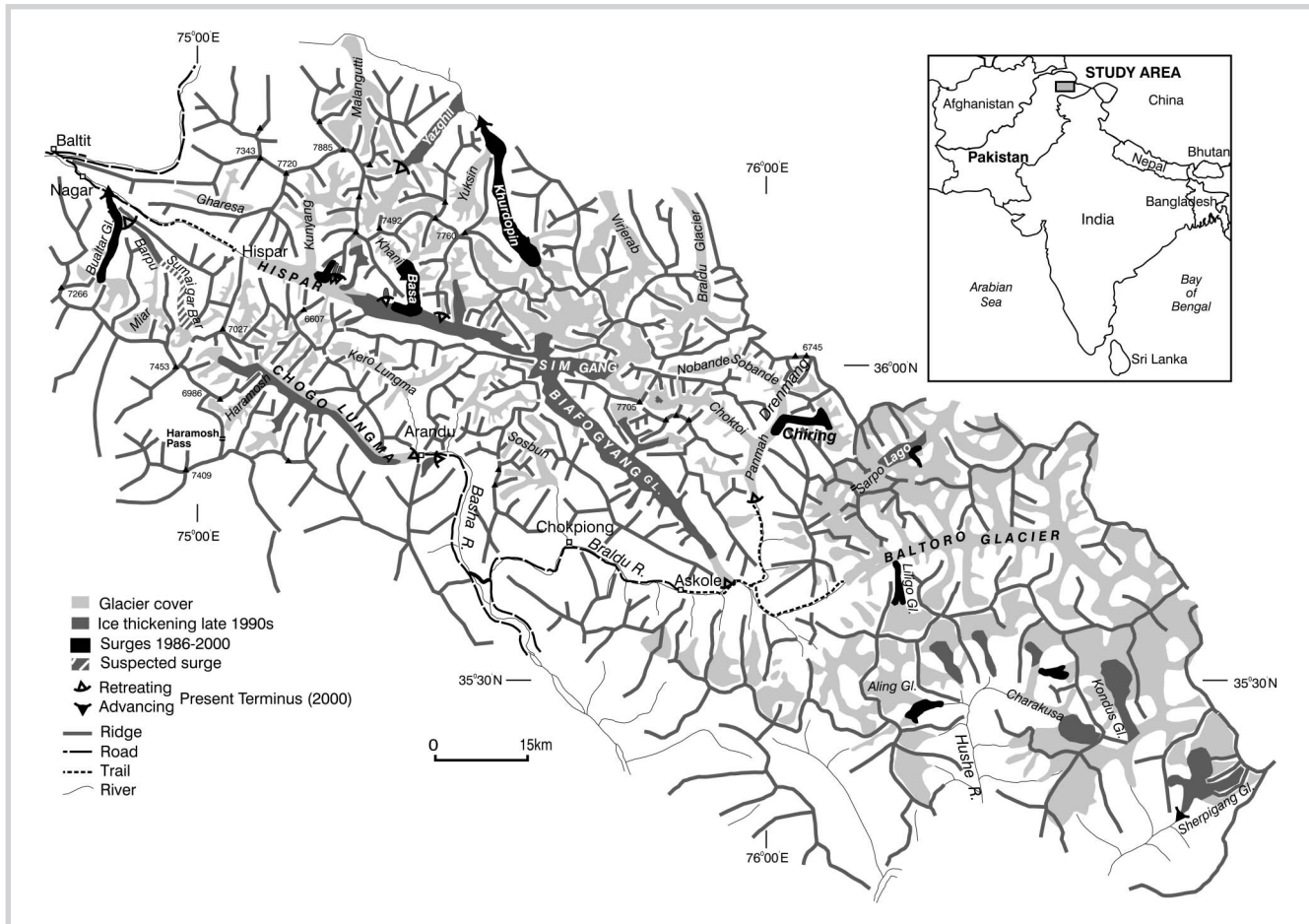
Biafo glaciers, comparing features and repeat photography at stations and transects where measurements were carried out between 1985 and 1989 (Hewitt et al 1989; Hewitt 1989). In a few cases the changes can be confirmed from satellite imagery. Specifically, LANDSAT imagery shows the steady advance of the Bualtar terminus between 1988 and 2000, as well as the 1987 surge; at Panmah Glacier the 2.5 km advance of Maedan tributary since the mid-1990s, the 1994 surge of Chiring, and continuing disturbance of the main glacier. The advance of the Aling, of the 2 tongues of the Yazghil, and the Khurdopin surge are also confirmed. However, most evidence for the expansions detailed here is not readily deduced from available satellite coverage. The ice-margin features, processes, and purely vertical shifts identified with thickening—or thinning—are too localized, or masked amid debris-covered ice of the termini and lateral margins.

The Shimshal glaciers, the Chogo Lungma, and the Sherpigang are included here because, although only visited once in 1999 or 2001, they exhibit features associated with expansion elsewhere. The recent advances were confirmed by local guides and, in the case of the Yazghil, by satellite imagery. An advance of the Sherpigang terminus of 1 and 4 m in different parts over a week in July 2001 was measured relative to proglacial markers.

Herein lies the importance of ground observations and the photographed examples (Figures 2 and 3).

They not only identify sites of recent expansion but distinctive ice margin geometries and associated processes. Responses to thickening at lateral margins in the ablation zones include:

1. Well-defined line of shear between active ice and passive margin.
2. Continuous sections of ice cliff rising steeply to the glacier surface, or over-riding and overhanging the glacier margins.
3. Fresh ridges of lateral moraine beside the line of shear (Figure 2), or
4. Supraglacial debris and ice blocks shed from overhanging ice onto the valley side flank of lateral moraines and into the so-called 'ablation valley' trough (Figure 3).
5. Reduction or closing of crevasses except on icefalls where they may increase (the main Sherpigang icefall, for example).
6. Closing of ice margin conduits to subglacial drainage and ponding of water at the ice edge.

Such features were found in all cases of glacier thickening identified here. Equally, they were absent at these sites a decade ago and rarely seen anywhere through the middle decades of the 20th century. Yet, they were commonplace in its first 2 decades and the

TABLE 1 Glacier changes observed in the central Karakoram, 1997–2001.

Glacier	Length (km)	Elevation: highest–lowest (m)	Aspect	Thickening, where observed (m)	Terminus advance
SHIMSHAL					
1. Khurdopin[a]	47	7760–3250	N	15–20 (lower glacier)	no
2. Yazghil	31	7852–3200	N	5+ (lower glacier)	yes (into river)
3. Malungutti	23	7885–2900	N	5–10 (lower glacier)	yes (into river)
HUNZA					
4. Pasu	25	7610–2550	E	2–5 (mid-glacier)	no
5. Ghulkin	19	7390–2450	E	5–10 (lower glacier)	yes (20 m+)
6. Hispar	62	7885–3000	W	10–20 (upper glacier)	no
7–11. Hispar tributaries			S	varied (lower glacier)	yes
12. Sumaiyar[b]	16	7027–3900	NW	10–15 (lower glacier)	yes (into Miar Glacier)
BRALDU					
13. Tippur	10	6400–2900	NE	10–15 (lower glacier)	yes
14. Biafo	68	7290–3150	SE	10–25 (mid-glacier)	no
15–18. Biafo tributaries			3NE, 1SW	varied (lower glacier)	yes
19. Maedan[c]	12	6200–4400	NW	10–40 m (lower glacier)	yes (2.5 km 1996–1999)
20. Unnamed[d]	9	6000–4300	SW	lower glacier	yes (1.5 km 1996–1999)
21. Baltoro[e]	60	8610–3490	W	?	yes (150 m 1990–1997)
HUSHE-KONDUS					
22. Sherpigang	20.5	7740–3520	SW	5–20 (all)	yes (2001)
23–28. Tributaries	—	—	3SE, 3NW	—	yes (2001)
29. Kaberi (Kondus)	26	7665–3300	S	5–10 (upper glacier)	no
30. Charakusa (Hushe)	21	7280–3480	W	10–20 (upper glacier)	no
31. Chupergyatrapsa[f]	3.5	6560–4440	W	10–30 (all)	yes (2001)
32. Ghondoghoro	19	7160–3420	S	10–20 (mid-glacier)	yes (2001)
33. Aling	20.5	6700–3660	SE	5–50 (all)	yes (1997–2001)

[a] may have surged
[b] tributary of Barpu Glacier
[c] tributary of Panmah Glacier
[d] valley immediately adjacent to Maedan, beside "Shinchapbi"
[e] Personal communication from Prof. C. Smiraglia, Milan, who finds Baltoro has advanced slowly since 1955
[f] detached tributary of Chogolisa (Charakusa)

late 19th century. They appear repeatedly in reports of the Conway, Workman, and de Filippi expeditions, for example. At that time they were assumed to reflect the distinctive mechanics of these glaciers, or so-called *blockschollen* movement. Rather, it seems they were characteristic responses in the Little Ice Age expansion.

By contrast, the thinning and retreat that dominate observations from the 1950s to the 1990s meant that debris slumping from moraines buried the line of shear and ice edge. The glacier margins at the sites where the recent changes occurred were previously subdued or chaotic, with open crevasses and stagnant ice. For some tens or hundreds of meters, irregular, debris-covered ice fringed the main ice streams. Stagnant ponds were common, often of clear water, and vegetation colonized the supraglacial moraine. Ice

margin and tributary streams plunged into large holes at the ice edge.

Similar contrasts apply between advancing and retreating termini. The former are generally steep-fronted, with extensive ice cliffs and active ice facets on the advancing lobes. Retreating termini have gentle fronts buried in debris that gradually blankets the whole to create a subdued, hummocky surface where ice is rarely seen and vegetation takes hold. However, the termini of large glaciers are insensitive or highly lagged indicators, hence the importance of the thickening observed up-valley of cases like Biafo and Barpu, whose termini continue to decline.

It should also be said that the difficulties of identifying most of these changes on satellite imagery apply equally to past and on-going thinning, retreat, and stagnation. The termini and lower 2–3 km of the Barpu, Hispar, Biafo, Panmah, Charakusa, and Kondus glaciers continued to waste back to 2001. Moreover, while indications of expansion of the higher and larger central Karakoram glaciers are the focus here, most glaciers to the north and west have continued to diminish. Except for 2 surges, a dozen glaciers revisited in 2000 or 2001 in the Upper Chapursan, Chalt, Naltar, Karambar, and Darkot valleys continued to thin and retreat. These changes cannot be deduced from comparisons of 1980s, 1990s, and 2000 LANDSAT imagery. Again, the problem for available technology is identifying purely vertical changes of less than 10–15 m, and heavy debris covers that tend to mask horizontal changes.

Meanwhile, the Karakoram involves distinctive glacial environments and features. In the past it has been suggested that the glaciers respond differently to climate changes due especially to the heavy debris covers on the ablation zones and to rare, large influxes of monsoon snowfall (von Wissmann 1959; Kick 1989). However, there is a need to recognize other conditions which, it is argued here, seem more likely to govern the sorts of changes seen recently.

Regional climate, seasonality, and recent climate change

At different times of year the glaciers come under the influence of 3 distinct weather systems. A westerly circulation and cyclonic storms dominate the winter. Measurements in the 1980s showed that two-thirds of high-altitude snow accumulation on central Karakoram glaciers occurs in winter (Hewitt et al 1989). The other one-third was summer snowfall, and suggests the monsoon may advance over the region in most years (Mayewski and Jeschke 1979; Wake 1989). The glaciers are intermediate between the 'summer accumulation' types of the Greater Himalaya, and 'winter accumulation' of the European Alps to the west (Ageta and Fujita 1996).


FIGURE 2   Biafo Glacier, mid-ablation zone, 1999. In the foreground 2 phases of a total of 15 m of thickening are observed. A fresh lateral moraine ridge (arrow) records recent passage of even higher ice. The glacier here is 3 km wide. The view is from above the Mango icefall at about 3800 m, towards Baintha Brakk (7385 m) some 35 km distant. (Photo by Kenneth Hewitt, July 1999)

FIGURE 3   The Sumaiyar Bar tributary of Barpu Glacier, Nagar, in 1999. Debris-covered ice overtops the highest lateral moraines where, in 1985–1989, the ice was 10–20 m below them. (Photo by Kenneth Hewitt, June 1999)

Continental anticyclones affect the behavior of the other 2 systems, including storm paths and incidence of clear weather. Since direct solar radiation is responsible for 80–85% of glacier ablation, clear-versus-cloudy weather in summer is a critical variable. Several relevant trends are reported in temperature and precipitation records at valley weather stations (Archer and Fowler 2004). Since the early 1960s, analyses show an increase in winter precipitation and declining summer mean and minimum temperatures. The latter is probably due to more summer storms, widely reported by local people, although no regular increase in summer precipita-

FIGURE 4 Vertical organization of the glacial zone relating area–altitude, ice and ice-free basin zones of two glaciers to temperature and precipitation. Barpu-Bualtar is a typical avalanche-nourished system. Biafo has large, open accumulation zone areas above 4800 m; the right-hand histogram shows the contribution of snowfall measured in different elevation zones. (Source: Hewitt et al 1989)
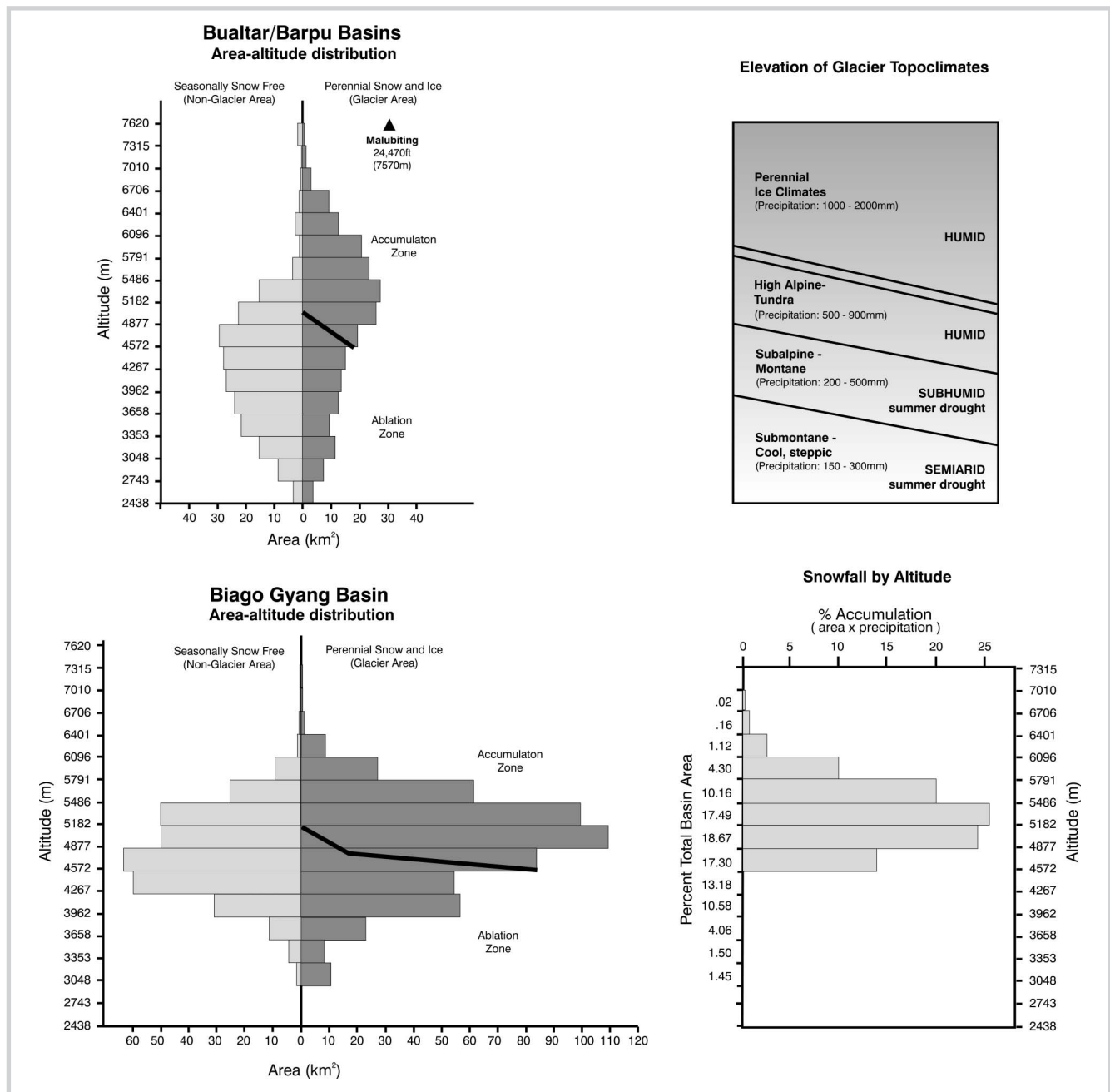
tion was measured. These developments suggest climate change consistent with positive mass balance on the glaciers. Increasing winter temperatures are in accord with warming effects worldwide, but less critical for these high-altitude glaciers. Much reduced flows are reported for rivers dominated by glacier runoff draining the highest parts of the Central Karakoram (Archer and Fowler 2004).

Unfortunately, the data come from stations far below the glacial zone and, while the trends may suggest positive mass balance, thinning and recession were

the rule through the mid-1990s, even in the highest basins. Why are the expansions confined to the latter while glaciers of intermediate elevations continue to decline? Why is there nothing in the climate records to indicate the suddenness of the expansions?

The 'elevation effect'

Glacier mass balance depends on relations between accumulation of snow on the upper glacier and ice ablation on the lower, moderated by flow conditions

between them. In the Karakoram there is a 5- to 10-fold increase in precipitation between glacier termini around 2500 m, and accumulation zones above 4800 m (Figure 4). Maximum precipitation occurs between 5000 and 6000 m. That is 2000–3000 m higher than is generally reported from the tropics (Barry 1992). A few glaciers like the Biafo have extensive, avalanche-free and large accumulation zones. Most others are nourished mainly or wholly by avalanched snow. As with the Barpu–Bualtar system, their ablation zones are larger than accumulation zones. In such glaciers, steep high rock walls acting like 'snow fences' trap more snow than their map area would suggest. Moreover, the extent of heavy supraglacial moraine, which greatly reduces ablation at lower elevations, is a direct function of avalanche nourishment.

Perhaps two-thirds or more of Karakoram ice derives from snow avalanched through at least 1000 m of elevation (Hewitt 1994). This allows conditions at higher altitudes to more quickly influence those downslope. Climate change may well alter the scale, pattern, and qualities of avalanches through high-altitude snowfall amounts, or summer versus winter snowfall. This is also a mechanism to alter ice temperatures and debris characteristics (see below).

Orographic effects on precipitation are decisive for glacier nourishment. Meanwhile, ablation is controlled by solar radiation and the seasonal migration of temperatures with elevation. Under favorable conditions ablation rates are high. On the Biafo, at 4080 m in July, measurements on bare or dusty ice recorded 50–90 mm of ice loss per day in clear or slightly cloudy conditions. Annual losses were in the range of 5–7 m (Hewitt et al 1989). However, huge differences occur under different ice surface conditions or due to topographic shading. On Batura Glacier at 2610 m in areas of thin supraglacial debris, annual ablation of 9 m was recorded, rising locally to 18.4 m. Yet heavy supraglacial moraine reduced average annual ablation to a few centimeters (Batura Investigations Group 1979). Even in the season and areas of highest ablation, day-to-day fluctuations are large and, over the larger part of ablation zones, melting lasts only a few weeks. Hence summer weather is critical.

Solar radiation causes most of the ablation, but temperatures and snow cover are major constraints. Ablation losses depend on temperatures above zero, and ice area exposed by melting of the seasonal snow cover. Each of these parameters declines in duration upwards to ablation zone limits around 5000 m. Summer storms obviously reduce sunshine and a snowfall shuts down ablation completely.

A crucial fact for mass balance and climatic response is that 60–80% of total ice area—most of the accumulation and ablation zones—lies between 3800 and 5800 m. The low penetration of the glacier termini is a distinctive feature, but it receives more attention than it warrants. In this regard, the important role of supraglacial debris needs to be reconsidered.

Supraglacial debris and dust

Heavy supraglacial debris or moraine covers the lower ablation zones of most Karakoram glaciers and is widely assumed to explain the low-altitude penetration of termini (Kick 1989). Debris thicker than 4–5 cm reduces ablation and it becomes negligible beneath 1–2 m. However, the extent and effect of supraglacial moraine has been exaggerated. More important generally, and more sensitive to secular climate change, is dusty and dirty ice (Figure 5). It involves much larger areas, specifically in the mid- to upper ablation zones above 3800 m where 70–80% of ice ablation occurs. During clear summer weather, seemingly bare or 'clean' ice becomes covered in wind-blown dust, thin sprinkles of sediment particles—the 'kryokonit' material of former interpretations—and films of algae. These increase rates of melting. Dirt layers a few millimeters thick can increase them by 40% or more (Mattson and Gardner 1989). Much of the seemingly continuous, heavy supraglacial moraine contains large areas of dirty ice, especially dirt-veneered ice facets that melt back at exceptionally high rates, offsetting the degree of protection by thicker debris (Khan 1989).

The ablation response of heavy versus light debris to climatic factors is opposite in sign and significance. Continuous heavy, ablation-suppressing debris is a passive and relatively conservative factor. Excepting glaciers that surge, or following catastrophic rockslides, its extent changes very little over centuries. Dust and dirt veneers are highly sensitive to summer conditions. In sunny weather the darker dust particles and dirt lead to pitting of the ice and a 'weathering layer' that adds to penetration and absorption of solar radiation. In cloudy weather dust and dirt tend to wash away, or snowfall covers them. When sunny weather returns there is a lag, often of several days, before melting of snowfall, build up of supraglacial dust, and the ice surface weathering layer restore high rates of ablation. The shortness of the main ablation season means small changes here can have large effects.

To summarize: the decisive altitudes for mass balance in the central Karakoram are roughly 4500–6000 m for net accumulation, and 3500–4300 m for ablation. The main zone where dust and dirty ice enhance ablation is 3500–4600 m, and for protection by supraglacial debris below 3500 m. The probable elevation of maximum ice flux is 4200–4500 m. This is somewhat below the ELA line of balance between net accumulation and net ablation for snowfall-nourished glaciers, owing to the importance of avalanche transfer and deposition.

FIGURE 5 Rates of ablation at monitored plots on Biafo Glacier at an elevation of 4100 m in 1986 illustrate the importance of debris cover. Types I through VIII each represent conditions over tens of square kilometers in the ablation zone as a whole. Thin debris promotes higher rates, shows larger fluctuations according to weather, wetting and drying, and sensitivity to small-scale patterns of build-up and disturbance of material (compare III, V and VII).

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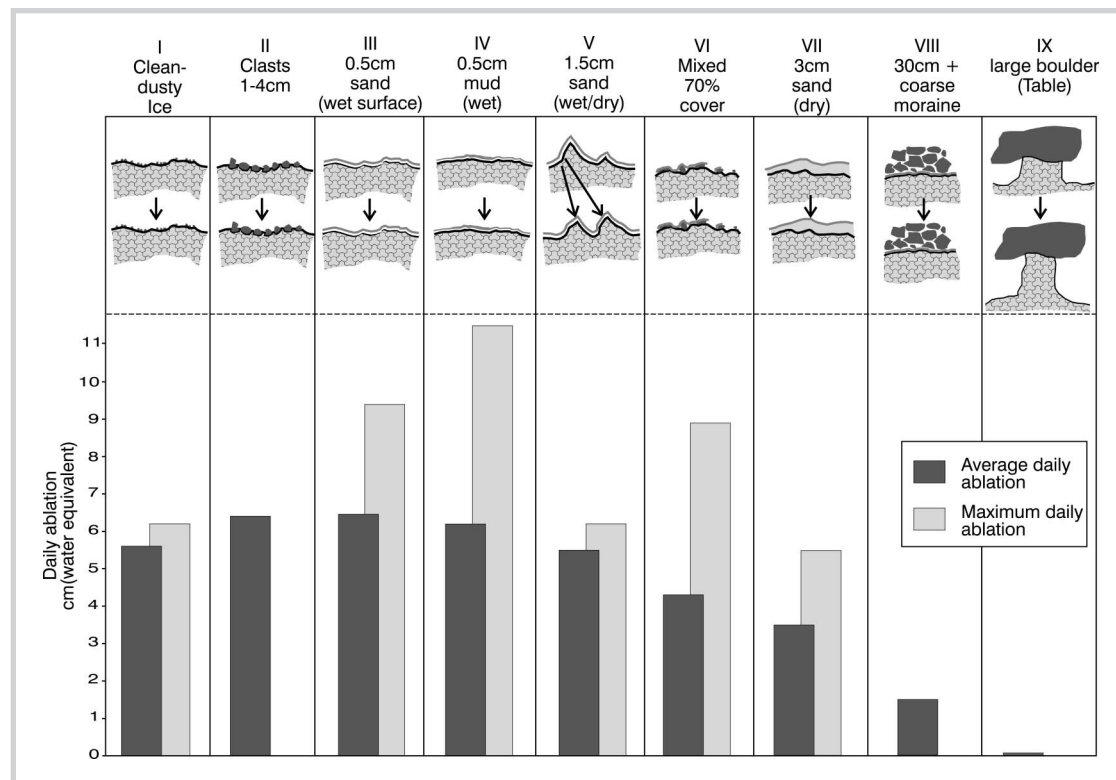
Combining these attributes shows the critical elevations for climatic responses lie between about 4000 and 6000 m. However, an indirect effect of verticality can further transform the impact of climate on glaciers.

Glacier thermal regimes

Paterson (1994: 205) recognizes 4 classes of vertical temperature distribution in glaciers:

1. "All ice is below melting point.
2. The melting point is reached only at the bed.
3. A basal layer of finite thickness is at the melting point.
4. All ice is at the melting point except for a surface layer, about 15 m thick, subject to seasonal variations in temperature."

The first 2 types are referred to as 'cold', the third as 'polythermal', and the fourth as 'temperate'. However, in high mountain glaciers all of these can occur, and even more complicated local profiles, according to elevation and orientation. Although there are few actual measurements of Karakoram ice, those available indicate that all 4 thermal classes can be present, often in different parts of single glaciers (Batura Investigations Group 1979; Hewitt 1989). The exceptional number of surging glaciers derives in part from these conditions (Hewitt 1969; Hewitt 1998a).

Ice temperature influences glacier behavior in 2 main ways: through the energy required to bring cold ice to the melting point, and glacier flow rates. The deformation rate of cold ice is temperature-dependent. It increases by a factor of 5 as ice is warmed from $-25°$ to $-10°$ C. Meanwhile, glaciers with cold ice tend to be frozen to the bed. Unfreezing can dramatically increase movement, sliding at the bed being a major component in warm ice and tending to increase with increased meltwater production. Paterson (1994: 337) recognizes the importance of these conditions by stating that: "Only in temperate glaciers is the effect of a climate change restricted to a change in mass balance."

In reality, however, ice temperature is largely ignored as a response to climate change, except long-term. Lag times for incorporation of warmer or colder snow, the critical factor, are generally very slow; the responses to ambient temperature or meltwater changes even more so. As it flows down valley, ice below the melting point is warmed slowly by friction near the bed and, to a small extent, by geothermal heating. The near-surface layers, and thin ice at the termini, undergo seasonal cooling in the winter freeze-up, but evidence elsewhere suggests lower elevation warmth penetrates so slowly below crevasse depth as to be insignificant. Yet subtropical, high relief glaciers involve factors that can accelerate these response times, especially compared to the mid-latitude and polar glaciers that have dominated relevant research. Certainly, because accumulation zone snowfall and avalanched snow are the primary determinants of ice temperature, in the Karakoram high-altitude climate and any changes occurring there will be decisive.

Is it possible the considerable climate changes since the late 19th century and, specifically, over the past 50 years, could alter thermal conditions sufficiently to help explain the Karakoram expansions? There are 3 distinct considerations: temperature change in high-altitude snowfall, its transfer to ice at depth, and sub-glacial areas close to the threshold between 'cold' and 'warm' ice in critical locations.

High-altitude snowfall

At lower elevations, reduced summer temperatures probably reflect greater cloudiness and/or increased precipitation. As well as reduced insolation, rain falling below the cloud layer and more moisture generally available for evaporation, should have a cooling effect.

However, at high altitude, increased cloud and precipitation increase thermal inputs through latent heat transfer. There is orographic forcing of condensation and advective transport of heat in the cloud layer. This will tend to raise snowfall temperatures; the more so if there is increased summer precipitation.

Unlike low-elevation rains, snowfall in these glacier accumulation zones largely occurs within the cloud layer. It is deposited at or close to cloud layer temperature, minimizing evaporative cooling. As noted, snowfall temperature largely determines glacier ice temperature, directly or through the thermal state of avalanched snow. However, is the 100+ year warming cycle since the Little Ice Age, even the past 50 years, enough to affect major glaciers and translate into increased ice flux lower down?

The highest relief glaciers appear to be divided into mainly cold, frozen bed ice in their upper reaches and warm, wet-based ice lower down. The situation is further complicated by warming of ice by surface melting, infiltration, and refreezing of moisture in summer. In the few glaciers with large, high-elevation firn basins and, according to orientation, surface ice may be warmer than at lower elevations. However, at Biafo and Hispar glaciers, the only cases with any measurements, sub-zero temperatures prevailed in firn profiles below 1 m depth, reflecting the predominance of winter snowfall (Hewitt 1989). This could change with greater summer snowfall but, with ice depths exceeding 400 m and increasing to 1100 m at their ELAs, it would take many centuries for such surface changes to affect thermal conditions down to and at the bed. In any case, the latter is most likely to occur first at pre-existing interfaces of different thermal regimes and have its strongest effects localized there, especially in steeply falling ice.

Avalanching, icefall regimes, and 'block flow'

Karakoram glaciers commence in steep tributaries, mostly icefalls and huge avalanche cones, that coalesce to form the main ice mass. There is all-year avalanching, but it is concentrated in spring and summer. Conversion to glacier ice and incorporation into these relatively thin and highly crevassed zones is much more concentrated in space, depth, and time—measured in years and decades rather than centuries. Most glaciers also have substantial icefalls between 4500 and 5500 m, and tributaries, often major ones, enter as extended icefalls where movement can exceed that elsewhere on the glacier as much as 10 times (Batura Investigations Group 1979). Icefalls are heavily and deeply crevassed. Much of the downslope movement occurs through unstable blocks and collapse of seracs. The ice thins considerably. Much more of the inner and subglacial parts are exposed to ambient weather conditions and meltwater influx.

The vertical 'cascade' between elevation zones, accelerated by steep ice, rapidly translates accumulation zone changes down glacier—although Oerlemans' (1989: 361) experiments suggest that icefalls, as opposed to generally steep ice, may delay or buffer the relation between mass balance change and responses of glacier termini. Critical locations for thermal regime change can coincide with, or be readily affected by, steep ice. This creates threshold conditions that can lead to abrupt change.

Concluding remarks

The anomalous expansions are confined to the highest relief glaciers and appeared suddenly and sporadically. This suggests the need to explore distinctive responses to high-altitude warming of snow and ice bringing thermal changes in thermally complex glaciers. Equally, we see the need to establish whether an increase in summer accumulation and decreased ablation in mid- to upper-ablation zone areas has brought a positive mass balance to the larger glaciers. What the Karakoram expansions do not do is refute the case for climate change, nor even atmospheric warming. The former seems the only explanation of the glacier changes observed. Warming and greater transport of moisture to higher altitudes may explain other aspects of the anomaly. The central Karakoram does emerge as the largest of those very few areas where glaciers are growing today, most probably due to the great elevations, relief, and distinctive climatic regimes involved.

Because 'disappearing' glaciers is the main prognosis for on-going climate change, Karakoram events may seem good news. After all, glacial meltwaters from the Karakoram dominate the flows of the main Indus and Yarkand rivers. As resources and hazards they directly affect the lives of some 200 million people in surrounding drylands. Unfortunately, no immediate benefit to thirsty lands downstream is assured. Declining river

flows from the highest central Karakoram watersheds over the past 2 decades suggest that any additional ice is going into long-term storage (Archer and Fowler 2004). Meanwhile, the more expanded glaciers in the 19th and early 20th centuries were associated with greater glacier hazards (Kreutzmann 1994; Hewitt 2004). If today's expansions mainly reflect redistribution of ice down-slope, eventually climatic warming could bring accelerated depletion. In this regard, the greatest problem is the absence of an established monitoring system for climate and other developments in the glacial zone, and relative neglect of this unique high mountain glacial region.

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