Glacier shrinkage and climatic change in the Russian Altai from the mid-20th century: An assessment using remote sensing and PRECIS regional climate model

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[1] This paper examines changes in the surface area of glaciers in the North and South Chuya Ridges, Altai Mountains in 1952–2004 and their links with regional climatic variations. The glacier surface areas for 2004 were derived from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) imagery. Data from the World Glacier Inventory (WGI) dating to 1952 and aerial photographs from 1952 were used to estimate the changes. 256 glaciers with a combined area of 253 ± 5.1 km^2 have been identified in the region in 2004. Estimation of changes in extent of 126 glaciers with the individual areas not less than 0.5 km^2 in 1952 revealed a 19.7 ± 5.8% reduction. The observed glacier retreat is primarily driven by an increase in summer temperatures since the 1980s when air temperatures were increasing at a rate of 0.10–0.13°C a^{-1} at the glacier tongue elevation. The regional climate projections for A2 and B2 CO_2 emission scenarios developed using PRECIS regional climate model indicate that summer temperatures will increase in the Altai in 2071–2100 by 6–7°C and 3–5°C respectively in comparison with 1961–1990 while annual precipitation will increase by 15% and 5%. The length of the ablation season will extend from June–August to the late April–early October. The projected increases in precipitation will not compensate for the projected warming and glaciers will continue to retreat in the 21st century under both B2 and A2 scenarios.

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1. Introduction

[2] Glaciers are one of the best indicators of climate change and a nearly global retreat of glaciers has been recently reported [*Kargel et al.*, 2005; *Barry*, 2006]. While glaciers of Europe and North America have been investigated in detail and there is a growing body of literature on the tropical glaciers, glaciers of Siberia remain underrepresented in the recent inventories. The climate of Siberia is characterised by extreme continentality expressed in strong seasonal temperature variations. Winter precipitation is meagre as a result of the very low winter temperatures and snow accumulation occurs between April and September at the same season as glacier ablation [*Dyurgerov and Meier*, 1999]. The response of such glaciers to climatic warming is understood less well then the response of glaciers with summer ablation and winter accumulation seasons. *Dyurgerov and Meier* [1999], *Fujita and Ageta* [2000], *De Smedt and Pattyn* [2003], and *Nakazawa and Fujita* [2006] have shown that mass balance of glaciers of the summer-accumulation type is strongly controlled by summer melt while the compensating effect of accumulation is limited. As a result, such glaciers may be more vulnerable to the observed climatic warming [*Fujita and Ageta*, 2000]. By contrast, a modelling study by *Braithwaite et al.* [2002] suggested that glaciers developing under a cold and dry continental climate with a short melt season are less sensitive to temperature variations. A further assessment of recent changes in the extent, mass balance, and climatic sensitivity of glaciers located in continental interiors is required.

[3] Of all Siberian mountains, the Altai Mountains (Figure 1) are most widely glaciated. Results of a glacier inventory in the Russian Altai based on the analysis of aerial photographs of the 1950s and field surveys of the 1960s were published in the various volumes of the Catalogue of Glaciers of the USSR (CG). According to the summary of the CG data by *Dolgushin and Osipova* [1989], glaciers covered 910 km^2 in the Russian Altai. The CG data included surface areas of individual glaciers, type of glacier, aspect, maximum and minimum elevations, and equilibrium line altitude (ELA) and are available as a part of the World Glacier Inventory (WGI) from the National Snow and Ice Data Center (NSIDC)

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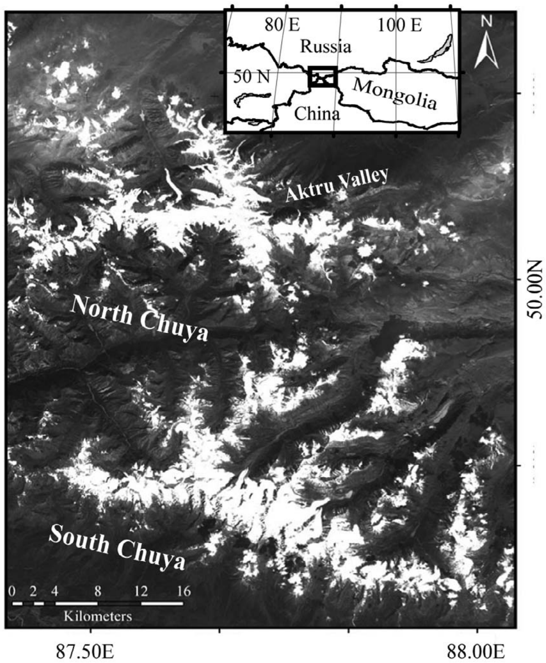

Figure 1. ASTER image from 10 September 2004 depicting the study area. Glaciers in the North and South Chuya Ridges are shown in white (the white colour represents clouds north and south of this area). The black rectangle on the inset map shows location of the Altai Mountains.

in Boulder, Colorado [*Bedford and Haggerty*, 1996; http://nsidc.org/]. Surface areas included clear ice and total areas (a combination of clear ice and debris-covered parts of glaciers). In the 2000s, investigations of area changes of small samples of glaciers were conducted in the Russian Altai [e.g. *Narozhny and Nikitin*, 2003; *Pattyn et al.*, 2003; *Surazakov et al.*, 2007] but an up-to-date regional assessment is lacking. An inventory including 91 glaciers was undertaken by *Kadota and Gombo* [2007] for the Mongolian Altai. These studies show that most glaciers have retreated since the mid-20th century, however, the reported retreat rates vary considerably between individual glaciers and regionally. In different regions of the Mongolian Altai, glaciers lost between 10% and 30% of their surface area since the middle of the 20th century [*Kadota and Gombo*, 2007]. In Russian Altai, the Sofiyskyi Glacier retreated twice as fast as the Malyi Aktru [*Pattyn et al.*, 2003]. Having investigated changes in the extent of seven glaciers in the Aktru valley in the last 50 years, *Surazakov et al.* [2007] commented on a strong impact of local factors on glacier change. The Altai glaciers repeatedly advanced and retreated throughout the Holocene and their latest widespread advance lasting for a few decades is dated to the 1800–1850s although there were shorter periods of advance in the early 20th century [*Solomina*, 1999]. Studies of the Holocene environments also point at often non-synchronous variations of glaciers located in close proximity. The uncertainty in evaluating rates of glacier change in the Altai introduced by local factors can only be reduced if a large sample including glaciers of different type and size classes is assessed using satellite glacier mapping.

[4] This paper has four objectives: (i) to estimate the surface area of glaciers in two most widely glaciated regions of the Russian Altai, the North Chyua and South Chuya Ridges, at the beginning of the 21st century; (ii) to evaluate changes in the glacier surface area between 1952 (the year of the previous large-scale glacier inventory) and the early 21st century using data published in *CG* [1974, 1977]; (iii) to analyse the observed glacier changes in the context of changes in air temperature and precipitation; (iv) to present the future regional climate change scenarios and discuss their possible implications for the Altai glaciers. The work was conducted in the framework of the Global Land Ice Monitoring from Space (GLIMS) project [*Bishop et al.*, 2004; *Kargel et al.*, 2005; *Raup et al.*, 2007] and used the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) imagery (Figure 1). Other glaciated regions of the Russian Altai, including the Katun Ridge and the easternmost sector of the South Chuya Ridge have not been included because of the cloud cover on the available ASTER images and snow cover on glacier tongues on the Landsat imagery from the 2000s.

2. Study Area

[5] The study region covers most of the North and South Chuya Ridges of the Russian Altai and extends between 50°15′–49°40′N and 87°15′–88°15′E (Figure 1). The elevations range mostly between 2100 m and 3800 m above sea level (a.s.l.) with the maximum elevations reaching 4183 m a.s.l. and 3960 m a.s.l. in the North and South Chuya Ridges respectively. Glaciers are located mainly above 2500 m a.s.l. This study examined 238 glaciers with a combined area of 335 km^2 (as in 1952). Of these, 53% where very small glaciers with an area less than 0.5 km^2 and 4.6% were larger than 5 km^2 [*CG*, 1974, 1977]. Four types, recognised by the WGI classification were distinguished including 56 valley glaciers with compound and simple basins (193 km^2 in the 1950s), 57 cirque (39 km^2), 70 hanging (20 km^2), 1 niche (0.1 km^2), and 1 crater (7 km^2) glaciers. In addition, two more classes were recognised by *CG* [1974, 1977] including 40 cirque-valley glaciers (69 km^2) and 13 flat-summit glaciers (7 km^2). The amount of precipitation decreases from the north-west to the south-east across the region and the equilibrium line altitude (ELA) increases approximately from 2600 m to 3600 m a.s.l. [*CG*, 1974, 1977]. The proportion of glacier surface covered by debris is comparatively low in the study region: 7% of the total glacier surface was covered by the supra-glacial debris in 1952 [*CG*, 1974, 1977].

[6] Seasonal cycles of air temperature and precipitation are extreme in the Altai. Located close to the centre of the Siberian high, the Altai is dominated by high atmospheric pressure blocking the westerly flow between November and March [*Panagiotopoulos et al.*, 2005]. The Siberian high develops in the lowest 1.5–2 km of the atmosphere reaching 3–4 km in the mountainous regions. The annual precipitation totals range between 470 mm and 540 mm at Kara-Tyrek, Akkem, and Aktru stations (Table 1; Figure 2) and the November–March precipitation accounts for 10–30% of the annual totals. The share of winter precipitation increases with elevation as the influence of the Siberian high diminishes.

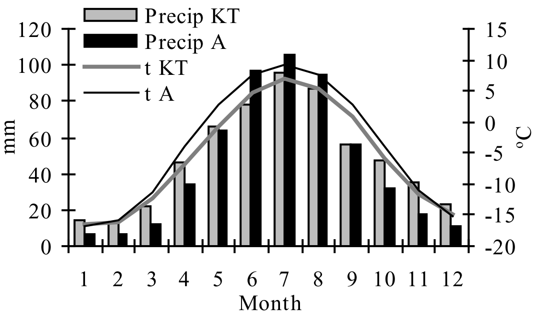

Figure 2. Monthly climatologies of air temperature and precipitation for the Akkem (A) and Kara-Tyurek (KT) meteorological stations for 1952–2004.

Between April and October, the westerly flow dominates. The North Atlantic depressions are the main source of precipitation accounting for 60% of annual accumulation while the role of depressions arriving from the south–west increases in autumn [*Aizen et al.*, 2006]. Snow accumulation occurs in the Altai throughout the year but predominantly in April–May and October when air temperatures are negative. The monthly air temperatures are positive at the station elevations in June–July–August (JJA) and ablation is limited to this period. Mass balance records from the Malyi Aktru Glacier, the World Glacier Monitoring Service (WGMS; http://www.wgms.ch) reference glacier for the Altai (50.08°N; 87.72°E; Figure 1) indicate that the average September–May and JJA mass balances were 680 mm and 780 mm of water equivalent (mm w. e.) respectively in 1975–2008 resulting in an average negative annual mass balance of 100 mm w. e. [*WGMS*, 2007 and earlier issues].

3. Data and Methods

3.1. Glacier Mapping

[7] The assessment of glacier surface area in the early 21st century and its changes since 1952 was conducted in three stages: (i) glaciers were mapped on the ASTER images and their surface areas for 2004 were calculated; (ii) accuracy of glacier surface area values published in *CG* [1974, 1977] was evaluated by re-mapping a sample of glaciers using the original aerial photographs; (iii) change in glacier surface area was calculated for the glaciers not smaller than 0.5 km^2 (as in 1952) selected from (i) and using data published in *CG* [1974, 1977]. For the glaciers that have split into several fragments, the net area change between 1952 and 2004 was based on a total area of the individual fragments.

3.1.1. Mapping Glaciers From ASTER 2004 Imagery

[8] Glacier outlines were mapped manually using ASTER imagery (resolution 15 m) from 10 September 2004 for the nearly cloud-free conditions (Figure 1). GLIMSView software (www.GLIMS.org) was used. The images were obtained from NASA Land Processes Distributed Active Center where they were orthorectified prior to the distribution using PCI OrthoEngine software [*Lang and Welch*, 1999]. The images were supplied in WGS1984 UTM zone N44 projection. The accuracy of the orthorectification was verified using a network of the interactive ground control points (GCP) established during the GPS ground surveys in August 2007. The hydrographic and topographic features on stable terrain in the vicinity of glacier termini were used as GCPs. The error of determination of GCP on the ground did not exceed 3–5 m and the root-mean-square-error (RSME$_{x,y}$) values were less than a size of ASTER pixel (15 m). Earlier assessments have confirmed that human interpretation remains the best tool for extracting detailed information from satellite imagery for glaciers if they are at least partly debris- and snow-covered particularly when mapping is conducted by the same person using a combination of different types of imagery [*Paul et al.*, 2002; *Raup et al.*, 2007]. The International Space Station (ISS) photographs taken on 9 August 2006 with a resolution of 5 m and an archive of terrestrial photographs were used for visual analysis providing additional information for the glacier boundary delineation.

[9] Snow, continuous debris cover and shadows are the well-known sources of uncertainty in glacier mapping. The extent of snow cover is at its annual minimum in the Altai at the end of August–beginning of September. On 10 September 2004, conditions were suitable for glacier mapping as glacier tongues were clearly visible on the image across the study area. The proportion of glacier surface covered by debris is comparatively low in the study region [*CG*, 1974, 1977]. Where glacier margins were obscured by shadows, a contrast-stretching function was applied to the ASTER imagery using ENVI 3.5 software. To estimate the error introduced by these factors, we compared areas of thirty randomly selected glaciers derived from the same ASTER image by three independent operators. This comparison yielded an error of ±2.1%. *Surazakov et al.* [2007] estimated areas of seven glaciers of the Aktru basin using high-resolution PRISM imagery for 2006. A comparison of these results with the areas of the same glaciers derived from ASTER 2004 imagery yielded an error of ±1%. An error of ±2.1% was, therefore, added to the 2004 glacier area estimation. Previous studies indicated a typical accuracy of glacier mapping of 3–4% for similar or larger samples of randomly selected glaciers [*Paul et al.*, 2002; *Bolch and Kamp*, 2006; *Kutuzov and Shahgedanova*, 2009].

3.1.2. Verification of the Catalogue of Glaciers Data Using Aerial Photographs

[10] The individual glacier surface areas published in *CG* [1974, 1977] for the study region were derived from the aerial photographs from 24 August–3 September 1952. The published values were verified by re-mapping a sample of 21 glaciers using a set of 37 original aerial photographs (Table 2). The aerial photographs were suitable for glacier mapping as glacier tongues were free of snow. The test glaciers chosen either among those surveyed in 2007 in the field or randomly ranged in size (0.8 km^2–22.6 km^2 as in 2004) and in changes of surface area.

[11] Resolution of the aerial photographs was 1–3 m. The paper prints of the aerial photographs were digitised at 600 dpi (1 m resolution). The instability of the paper base introduced distortions of 0.2–0.6 pixel (about 0.2–1.7 m in

Table 1. Meteorological Stations Used in This Study

Meteorological station	Lat (N)	Lon (E)	Altitude (m)	Observation period
Kara-Tyurek	49° 59′	86° 28′	2600	Since 1939
Akkem	49° 55′	86° 32′	2056	Since 1950
Aktru	50° 05′	87° 47′	2110	1958–1994

Table 2. Assessment of Surface Area Calculation and Changes in Surface Areas of Test Glaciers Using *CG* [1974, 1977; A_{cat}], Aerial Photographs for 1952 (A_{1952}), and ASTER Imagery for 2004 (A_{2004})[a]

Glacier	A_{cat} (km^2)	A_{1952} (km^2)	$A_{cat} - A_{1952}$ (km^2)	$A_{cat} - A_{1952}$ (% A_{cat})	A_{2004} (km^2)	ΔA_{cat} (%)	ΔA_{1952} (%)
Bolshoi Abyl-Oyuk	5.0	4.45	0.05	11	4.19	16.2	5.8
Bolshoi Maashei	16.0	15.95	0.05	0.3	14.03	12.3	12
Bolshoi Taldurinskyi	28.2	25.81	2.39	8.5	22.6	19.9	12.4
Jello	8.5	7.84	0.66	7.8	7.26	14.6	7.4
Kurumdu	5.2	5.02	0.18	3.5	4.58	11.9	8.8
Kurkurek	3.2	2.98	0.22	6.8	2.32	27.5	22.1
Levyi Aktru	6.5	6.15	0.35	5.3	5.84	10.2	5.0
Levyi Karagemskyi	6.4	6.45	−0.05	0.1	6.11	4.5	5.3
Nekrasov	2.6	2.47	0.13	5.0	1.74	33.1	29.6
Pravyi Aktru	4.8	5.27	−0.47	−9.7	4.62	3.8	12.3
Pravyi Karagemskyi	2.8	2.47	0.33	11.8	2.29	18.2	7.3
Sofiyskyi	17.6	17.45	0.15	0.8	14.93	15.2	14.4
Udachnyi	4.8	4.7	0.10	2.0	4.18	12.9	11.1
Shavla	3.4	3.25	0.15	4.4	2.71	20.3	16.6
Yadrintsev	8.0	7.75	0.25	3.1	6.04	24.5	22.1
SU5A15105005	1.4	1.28	0.12	8.6	1.24	11.4	3.1
SU5A15105024	1.0	1.04	−0.04	−4.0	0.95	5.0	8.6
SU5A15105171	1.0	1.0	0	0	0.94	6.0	6.0
SU5A15105234	1.3	1.36	−0.06	−4.6	1.19	8.5	12.5
SU5A15106071	1.5	1.35	0.15	10	1.18	21	12.6
SU5A15106094	1.2	1.1	0.1	9.1	0.87	27.5	20.9
Total/**Average**	130.4 ± 3.39	125.14 ± 4.83	4.76	**5.5**	109.81 ± 2.2	15.5 ± 0.51	12.2 ± 0.53

[a]Changes in glacier area between 2004 and 1952 are according to CG (ΔA_{cat}) and results of re-mapping (ΔA_{1952}). Absolute values of $A_{cat} - A_{1952}$ (% A_{cat}) are used to calculate the average difference (shown in bold).

ground distances of the digitised images). The digitized aerial photographs were co-registered to the orthorectified ASTER image using a network of 10–15 GCP established during the ground surveys per individual photograph. This procedure was carried out using ERDAS Imagine 9.0 software. After co-registration, the $RSME_{x,y}$ values not exceeding 7 m were achieved which is less than a size of ASTER pixel. Following co-registration, ice margins in glacier ablation zones were manually derived from the aerial photographs. Glacier margins in the accumulation zones, where glaciers have not retreated, were delineated from the 2004 polygons derived from ASTER imagery to avoid a well-known problem associated with a very low contrast on the snow-covered upper parts of glaciers on aerial photographs. The ice divides are clearly expressed in the study area and for most selected glaciers these were determined in the previous field surveys. The 1: 50,000 maps from 1958–1965 were used in the delineation of ice divides for the compilation of *CG* [1974, 1977].

[12] The accuracy of the re-calculated areas of the 21 test glaciers depends on the errors of (i) co-registration of aerial photographs and ASTER image and (ii) delineation of glaciers margins introduced by snow, debris cover and shadows. The error of co-registration was estimated by calculating the area of a buffer created around each test glacier [*Granshaw and Fountain*, 2006; *Bolch et al.*, 2010]. The width of the buffer was chosen as the maximum value of $RSME_{x,y}$ (7 m) resulting in an error of ±3.47%. An error of manual delineation was estimated through repeated measurements of the outlines of 21 test glaciers by three independent operators as ±1.7%. The total error calculated as root sum square of the two error terms was ±3.86%.

[13] The accuracy of the 1952 glacier areas published in *CG* [1974, 1977] is difficult to evaluate as these data exist as tables. The CG provides error terms for different size classes of glaciers but does not comment on other error types. The error term derived from CG has been assigned to the CG area of each test glacier. An additional error of 5% (based on the independent mapping of a sample of debris-covered glaciers tongues and considerable vertical stretches) has been added to the areas of such glaciers to account for a more difficult delineation. The resulting average error for the 21 test glaciers was ±2.6%.

[14] The derived areas of the test glaciers were compared with those listed in CG (Table 2). The average difference was 5.5% of the CG values. This is close to a difference of 5% obtained in a similar comparison by *Bolch and Marchenko* [2009] for the Zailiyskyi Alatau. The Ryan-Joiner test has shown that the differences are normally distributed. The 5.5% difference is outside the uncertainty associated with the two samples although the ±2.6% error of the CG values is likely to be underestimated. The values published in *CG* [1974, 1977] mostly exceed those obtained by re-mapping (Table 2). The largest differences characterized Bolshoi Abyl-Oyuk (11%) and Pravyi Karagemskyi (12%) Glaciers. The extensive rock outcrops on Bolshoi Abyl-Oyuk were not excluded from the calculation of its surface area in the CG resulting in an overestimation of its published value. By contrast, the re-calculated area of the Pravyi Aktru Glacier exceeded that published in CG by nearly 10%. The most likely source of difference is a comparatively large extent of debris cover on Parvyi Aktru.

[15] The combined surface area of the test glaciers declined between 1952 and 2004 by 15.4% and 12.2% if the *CG* [1974, 1977] and the re-mapped 1952 values are used respectively. Therefore, 5.5% difference between the two 1952 data sets results in 3.3% difference in the calculation of the 1952–2004 glacier area change for a sample of 21 glaciers. This difference is within the uncertainty estimated as the root sum square of the total 1952 and 2004 errors and equal to ±3.28% if the CG data are used and ±4.34% if the re-mapped values are used. The difference is significantly smaller than the

observed shrinkage rates. Therefore, the *CG* [1974, 1977] data can be used to assess changes in glacier areas in the study region. The 5.5% uncertainty value in the 1952 data appears to be more realistic than ±2.6% error derived from the CG assessments of accuracy based on glacier size alone. The ±5.5% error was assigned to the 1952 CG data.

[16] Glaciers smaller than 0.5 km^2 (as in 1952) have been excluded from the assessment of changes because the *CG* [1974, 1977] reports a large error in estimation of glacier size for the very small glaciers increasing from about 5% for 0.5–1 km^2 glaciers to 10–25% for glaciers under 0.5 km^2. A limited availability of the original aerial photographs did not allow re-mapping of a sufficient sample of very small glaciers to verify accuracy of the reported values. Changes in the extent of 126 glaciers, whose individual glaciers were not less than 0.5 km^2 in 1952, have been evaluated in this study although a potentially significant contribution of very small glaciers to the net reduction in glaciated area is acknowledged [e.g. *Paul et al.*, 2004].

3.2. Climate

[17] Two conventional meteorological stations, Akkem and Kara-Tyurek located in the Katun Ridge, supply data for the middle mountains in the Altai (Table 1). Glacier termini descend to approximately 2300–2600 m a.s.l., which is close to the station elevation. The Aktru station located in the North Chuya Ridge (Figure 1) was closed in 1994. The JJA temperature averages at the three stations correlated closely with correlation coefficients of 0.81–0.96. The correlation coefficients between annual precipitation totals were lower (0.56 and 0.42 between Kara-Tyurek - Aktru and Akkem – Aktru respectively) but statistically significant at 0.05 confidence level. The time series of air temperature and precipitation from Akkem and Kara-Tyurek were used to evaluate the observed variations in regional climate. The starting year for precipitation time series was taken as 1952 when Tretyakov gauge replaced Nipher gauge. Uncertainties exist in gauge measurements of precipitation in Siberia due to (i) wetting and evaporation losses from the gauges, (ii) underestimation of trace precipitation, and (iii) wind-induced gauge undercatch of precipitation [*Groisman and Rankova*, 2001]. The wetting loss correction was implemented in the precipitation records prior to archiving the data by the USSR/Russian Hydrometeorological Service but other corrections were not. The evaporation losses from the Tretyakov gauge are not considered a source of significant bias in cold regions. Previous analysis of trace precipitation measurements at 61 stations in Siberia has shown that it results in an underestimation of annual precipitation by 1–12% being more important in the months and regions of low precipitation. Thus, at the Barnaul station located in the foothills of the Altai and characterised by lower precipitation amounts than Akkem and Kara-Tyurek, the largest underestimation of precipitation was about 8% in March and 3% in July and close to zero in other months. The wind-induced gauge undercatch introduces a more significant bias especially for snow. It was not possible to calculate corrections for wind-induced undercatch for the Akkem and Kara-Tyurek precipitation because the required daily wind speeds at gauge level, daily maximum air temperatures, and type of precipitation (rain, sleet or snow) were not available for this study. The daily data from a Campbell Scientific automatic weather station (AWS) for October 2007–April 2008 were used to approximately evaluate the catch ratio of snow (amount caught by the gauge) in the Altai using equation (1) after:

$$\mathrm{CR} = 103.10 - 8.76\,\mathrm{w} + 0.30\,\mathrm{t}_{\max} \qquad (1)$$

where CR (%) is catch ratio, w is daily wind speed (m s^{-1}), and t_{max} is daily maximum air temperature (°C). The AWS was installed at 2380 m a.s.l. (between elevations of Akkem and Kara-Tyurek stations; Table 1) near the tongues of the Aktru glaciers (Figure 1) and measured air temperature and wind speed every 5 seconds recording the average values every 15 minutes. The anemometer was installed at 2 m above the surface, i.e. at the same elevation as Tretyakov gauges at the conventional stations. To avoid possible overcorrection caused by blowing snow, days with mean daily wind speeds above 6.5 m s^{-1} [a threshold for blowing snow events] were excluded. These constituted only 2.5% of all days due to the domination of the Siberian high between November and March confirming conclusion that blowing snow events have limited impact on precipitation measurements in southern Siberia. The catch ratio varied between 80–81% for the November–March period and 82–84% for October and April–May.

[18] Regional climate change scenarios were generated using PREdicting Climate for Impact Studies (PRECIS) regional climate modelling (RCM) system developed by the UK Met Office as a user-friendly version of HadRM3 RCM [*Jones et al.*, 2004]. PRECIS is a hydrostatic, primitive-equation model with a horizontal resolution of 25 km. Fractional grid box land cover is used to improve detail of surface characterisation. PRECIS derives lateral boundary conditions from HadAM3P, a global atmosphere-only model with a resolution of 150 km, which is in turn forced by surface boundary conditions from the global circulation model HadCM3. PRECIS (HadRM3) has been successfully used in the mountains before [e.g. *Frei et al.*, 2003; *Schmidli et al.*, 2007; *Shahgedanova et al.*, 2009]. Three integrations have been performed for two time slices: (i) 1961–1990 providing 'baseline' climate and (ii) 2071–2100 providing the future climate change projections for the aggressive A2 and moderate B2 CO_2 emission scenarios assuming an increase in mean CO_2 concentrations to 830 ppm and 600 ppm by 2100 respectively [*SRES*, 2000]. The model domain extended between 40.5–61.5°N and 67.5–130.5°E. Model output for an area between 48–50.5°N and 86–89°E encompassing the Altai Mountains was used.

4. Results

4.1. Glacier Mapping

[19] Analysis of the ASTER imagery has shown that in 2004 there were 256 glaciers in the study region with a combined area of 253 ± 5.1 km^2. This represents a net surface area reduction of 82 km^2 (from 335 km^2 as in 1952 including glaciers smaller than 0.5 km^2). In 1952, there were 238 glaciers in place of the current 256. Of these 238, 222 glaciers have been identified on the ASTER imagery. Another 16 glaciers have fragmented and now constitute 34 glaciers. In 2004, 137 glaciers had an area below 0.5 km^2; 87 were

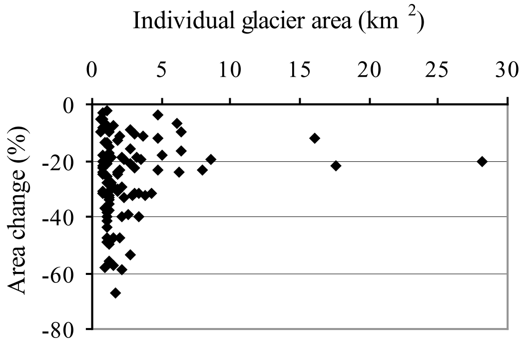

Figure 3. Change in glacier area between 1952 and 2004 versus glacier size (as in 1952).

between 0.5 km^2 and 2 km^2, 23 were between 2 km^2 and 5 km^2, and 9 exceeded 5 km^2. Analysis of changes in 126 glaciers with areas no less than 0.5 km^2 (as in 1952) showed that their combined area declined from 284 ± 15.6 km^2 in 1952 to 228 ± 4.6 km^2 in 2004, which constitutes a 19.7 ± 5.8% reduction.

[20] Glaciers of different size classes have lost approximately the same combined area but glaciers smaller than 2 km^2 have lost a much greater proportion of their areas (Table 3). On average, glaciers smaller than 2 km^2 have lost 28% of their area but changes in their surface area exhibit a wide range of variability from 2% to 68% (Figure 3) similarly to other regions, e.g. the Swiss Alps [*Paul et al.*, 2004], northern and central Tien Shan [*Bolch*, 2007; *Kutuzov and Shahgedanova*, 2009]. In 1952, glaciers with the individual areas of 0.5–1 km^2 accounted for 16% of the glaciated area excluding glaciers smaller than 0.5 km^2 (Table 3) and in 2004, they accounted for 23% of the decline in glaciated area. Glaciers larger than 5 km^2 accounted for 38% of the area and 32% of the change. The uncertainty in the estimation of 1952 areas of the glaciers larger than 5 km^2 was lower (Table 2) making it possible to discuss changes in individual glaciers. These glaciers lost between 4.5% (Levyi Karagemskyi) and 24.5% (Yadrintsev) of their 1952 areas. One of the largest in the Russian Altai, Bolshoi Taldurinskyi Glacier had an area of 28 km^2 in 1952 (26 km^2 according to the re-mapping results; Table 2) but has now separated into two glaciers with the combined area of 22.6 km^2 (Figure 4a). Bolshoi Maashei Glacier (6.1 km^2 in 1952) is close to fragmentation into at least three segments (Figure 4b). Sofiyskyi Glacier (17.6 km^2 in 1952 and 14.9 km^2 in 2004) has lost 2.7 km^2 or 15% of its area and its eastern tributary is close to separation from the main body of the glacier (Figure 4c).

[21] Glaciers with different aspects lost approximately the same area with an exception of east- and west-facing glaciers, which have lost 43% (highest loss) and 12% (lowest loss) of their 1952 areas respectively. In 1952, the west- and east-facing glaciers had the same average size of 1.3 km^2 and close (1.2 km^2 and 1 km^2) median values. The difference in relative loss is not an artifact of glacier size but might have resulted either from a small number of glaciers in the sub-sample (5 west-facing and 17 east-facing glaciers with a size over 0.5 km^2) or from specific local conditions. The west-facing glaciers are few in the study area due to snow drift; they develop in the topographically favorable structures and are less sensitive to climatic variations. While a map of area changes is not presented due to the potentially large errors in assessment of shrinkage of individual glaciers (Table 2), visual examination of the geographical patterns of both absolute and relative area changes suggested that glaciers located on the southern slope of the South Chuya Ridge experienced the strongest shrinkage possibly due to the higher incoming solar radiation and lower precipitation [*CG*, 1974, 1977]. Glacier size is another factor: the average

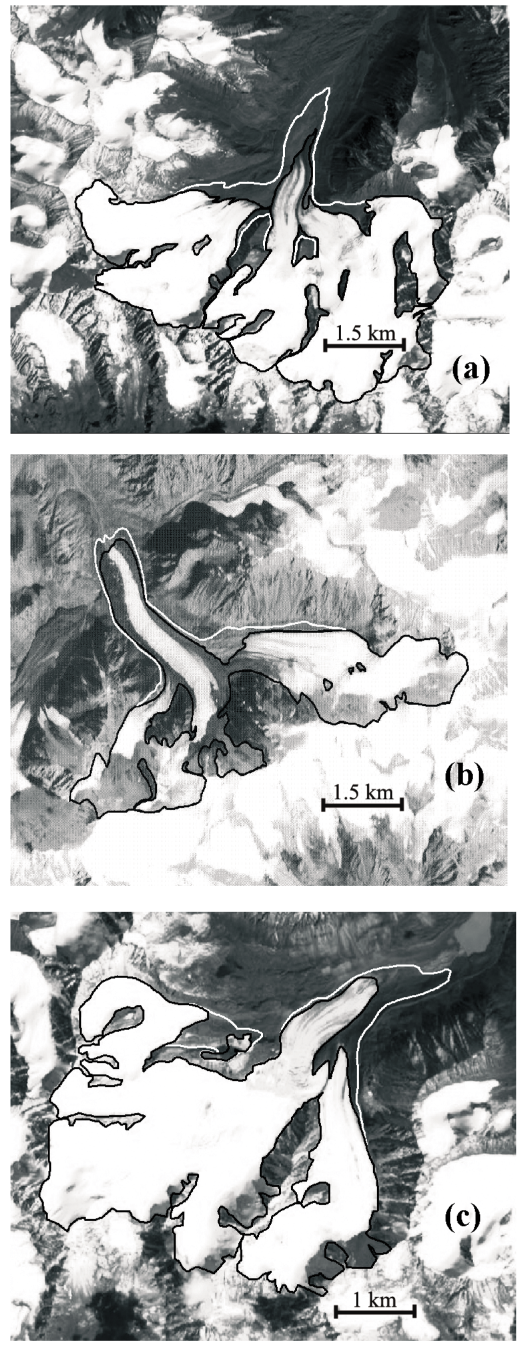

Figure 4. Retreat of the (a) Bolshoi Taldurinkyi, (b) Bolshoi Maashei, and (c) Sofiyskyi Glaciers located the South Chuya Ridge between 1952 (white line) and 2004 (black line). Aerial photographs from 1952 are used as background.

Table 3. Reduction in the Combined Area of Glaciers (km^2) as a Function of Glacier Size Class

Glacier size	Combined area in 1952	Combined area in 2004	Area loss (km^2)	Area loss (%)/st. deviation
>5	108.7	90.7	18.1	16.7/6.9
5-2	79.0	68.0	11.1	14.0/14.6
1–2	52.5	38.3	14.2	27.1/16.4
0.5–1	45.1	32.5	12.6	27.8/16.1

size of the examined glaciers is 1.4 km^2 on the southern slopes of both ridges and 1.9 km^2 and 2.0 km^2 on the northern slopes of the North and South Chuya Ridges respectively.

[22] The valley glaciers exhibited the largest absolute combined reduction in surface area (Table 4). The highest average loss per glacier characterized cirque-valley glaciers shrinking at an average rate of 0.08 km^2 a^{-1} per glacier. They also exhibited the largest relative area loss. It is the tongues of the cirque-valley glaciers that retreated rapidly while sectors resting in cirques experienced less change. The cirque glaciers lost smaller proportion of their surface area as their regime is strongly controlled by shading, snow drift and avalanche nourishment allowing them to survive under the climatic warming and even below the mean climatic ELA [*Kuhn*, 1995].

4.2. The Observed Climate Variations and Future Climate Change Scenarios

[23] The positive trends in JJA air temperatures between the mid-20th century and 1994 particularly below 2000 m elevation has been reported by *Narozhny and Adamenko* [2000]. At the Akkem station (2056 m a.s.l.; Figure 5a), a positive linear trend explained 25% of the total variance in the record in 1950–2004 yielding a temperature increase of 1.26°C. Following the negative anomalies observed in the 1980s, the warming intensified between 1985 and 2004 with the JJA temperatures increasing at a rate of 0.10°C a^{-1} and linear trend explaining 61% of the total variance. At Kara-Tyurek (2600 m a.s.l.), the long-term trend was weaker yielding a temperature increase of 0.87°C in 1940–2004 but in the last twenty years the JJA temperatures were increasing at 0.13°C a^{-1} (Figure 5c). At Kara-Tyurek, annual precipitation increased slightly since the 1950s with a linear trend explaining 14% of the total variance but at Akkem annual precipitation exhibited considerable variability but no significant linear trend (Figures 6b, 6d).

[24] Prior to the interpretation of regional climate change scenarios, model validation has been performed using data from five meteorological stations located between 48–50.5°N and 86–89°E for the 1961–1990 period. The gridded observational data sets have no advantage over individual stations for the purpose of model validation in the study area because there are few high-altitude stations. The average altitudes of model validation domain and of the stations were 2240 m and 2140 m a.m.s.l. respectively. The observed and modelled air temperatures are in close agreement throughout the year. In summer, the model overestimates monthly temperatures by 0.6–0.9°C (Figure 6a) possibly due to the under-representation of local convection and cloud cover.

[25] Precipitation intensity is most relevant for snow accumulation but it is one of the most challenging variables for simulation by both RCM and statistical downscaling [*Schmidli et al.*, 2007]. The model reproduces annual cycle of precipitation intensity but modelled values exhibit positive bias between October and May while the July–September precipitation is underestimated (Figure 6b). Between November and March, the modelled values of precipitation intensity exceed the observed by 13% in January and 30–33% in other months. In the months when most of snow accumulation occurs, this difference is lower: 17–26% in April–May, and 9% in October. There are two likely sources of this discrepancy: (i) undercatch of precipitation by the Tretyakov gauges and (ii) overestimation of the westerly flow and consequently precipitation by the model. An average wind-induced undercatch of precipitation between November and March is 19–20% and in October, April, and May is 16–18% at the elevation of 2380 m a.s.l.. The level of uncertainty implied by the use of uncorrected observed precipitation is close to the discrepancy between the observed and modelled precipitation intensity in cold season. A secondary source of discrepancy is a negative bias characterising the modelled sea level pressure (SLP) in the study area. A comparison between the 30-year (1961–1990) averages of December–February (DJF) SLP from the PRECIS simulation and from ERA 40 reanalysis [*Uppala et al.*, 2005] has shown that while the Siberian high, that dominates northern Asia in winter is reproduced by the model, the modelled high pressure area is shifted towards north–east. As a result, the 1961–1990 average of the DJF SLP is underestimated by the model by 7–10 hPa in the south–western part of the modelling domain extending to eastern Kazakhstan and the western Altai. This negative bias in the modelled SLP results in an overestimation of the westerly flow over the Altai leading to an overestimation of winter precipitation. Similar biases in RCM simulations of SLP have been reported by *Giorgi et al.* [2004] for the Eurasian continental areas. The bias in the simulated SLP is lower in January than in November–December and February–March. As a result, the difference between the modelled and the observed precipitation intensities in January is the lowest in the cold season and is very possibly due to the precipitation undercatch entirely. The discrepancy between the observed and modelled precipitation in October, April and May, when the atmospheric circulation is not affected by the

Table 4. Area Reduction for Different Types of Glaciers (With Individual Areas not Less Than 0.5 km^2 in 1952)[a]

Glacier type	Number	Average size, km^2	Area loss, km^2	Area loss, %
Valley	56	2.8	34.8	18.2
Cirque-valley	40	1.1	16.9	27.7
Cirque	18	0.8	2.7	16.3
Flat-summit	4	0.9	0.3	7.4
Hanging	7	0.6	0.7	14.4
Crater	1	6.4	0.6	9.6
Total/average	126	1.8	56.0	19.7

[a] The average sizes of glaciers refer to 2004.

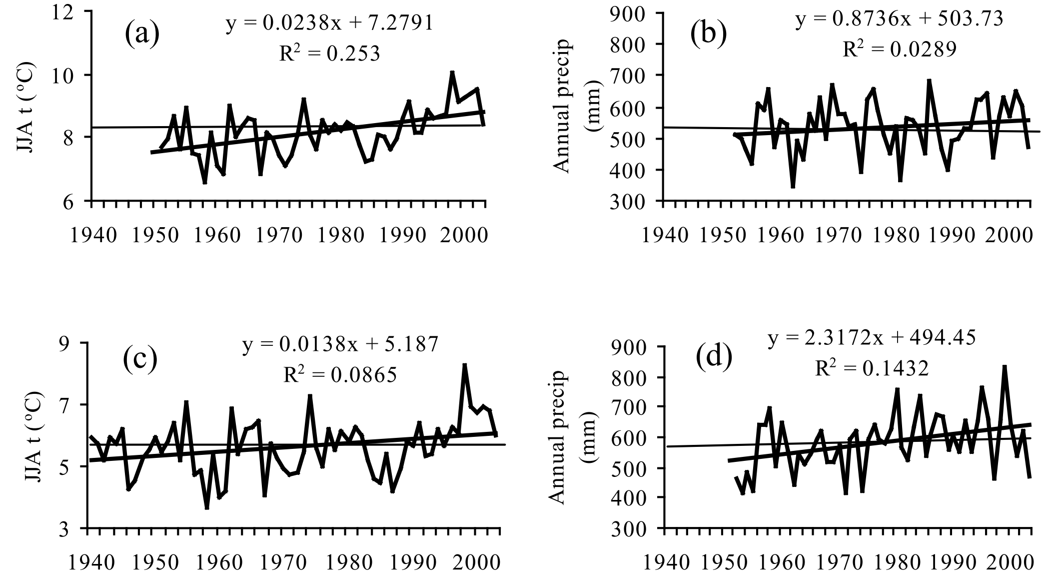

Figure 5. Time series of JJA air temperature and annual precipitation for Akkem (a, b) and Kara-Tyurek (c, d) stations. A straight bold line shows linear trend and a thin line shows record average.

Siberian high, can be attributed to the gauge undercatch of precipitation. Similar biases due to the uncertainties in precipitation measurements in cold and transitional seasons were reported for Eurasia by *Giorgi et al.* [2004] and for the European Alps by *Frei et al.* [2003]. A trend towards drier than observed conditions in summer is due to the underestimation of mesoscale convective precipitation by the model [*Frei et al.*, 2003; *Schmidli et al.*, 2007]. *Frei et al.* [2003] reported a negative bias of 35–40% in August–September precipitation in the Alps and similar differences are observed in the Altai simulation.

[26] PRECIS is able to reproduce the annual precipitation cycle accurately. The high precipitation intensity of PRECIS in the cold season is very realistic when a bias in the gauge-measured precipitation is taken into account. In particular, in April–May and October, the most important months for snow accumulation on the Altai glaciers, the difference between in the modelled and observed data (Figure 6b) is accounted for by the precipitation undercatch and this compares favour-

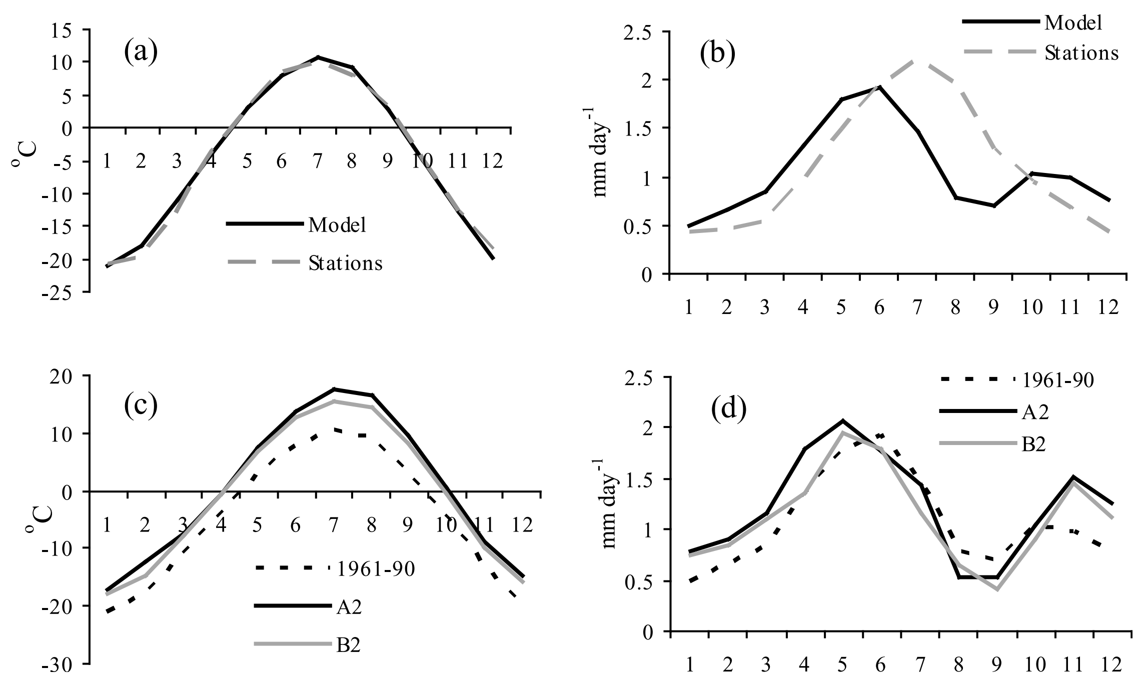

Figure 6. Modelled versus observed mean monthly temperatures (a) and precipitation intensity (b) for 1961–1990 and regional air temperature (c) and precipitation intensity (d) scenarios for the Altai Mountains (48–50.5°N; 86–89°E).

ably with RCM performance in the Alps [*Frei et al.*, 2003; *Schmidli et al.*, 2007] and in the Caucasus [*Shahgedanova et al.*, 2009]. The July–August precipitation underestimated by the model has limited impact on accumulation in the study region. Snowfalls do not occur on the Aktru glaciers when air temperatures exceed 5°C at the Kara-Tyurek station [*Narozhny and Adamenko*, 2000] and these have been consistently above the threshold after the 1980s (Figure 5c). Therefore, the developed projections can be used in assessment of changes in annual precipitation affecting accumulation between the baseline period and 2071–2100. PRECIS integrations for 2071–2100 indicate that JJA air temperatures will increase by 6–7°C and 3–5°C under the A2 and B2 group of SRES CO_2 emission scenarios in comparison with the baseline period of 1961–1990 (Figure 6c). Annual precipitation is projected to increase by 15% and 5% respectively (Figure 6d) with the largest increase in November–December in line with the projected weakening of the Siberian high [*Panagiotopoulos et al.*, 2005]. Importantly, the average air temperature in May and September will increase from 2.7–3.0°C to 6.7–7.5°C and in April and October from −3.8°C and −4.7°C respectively to −0.4–+0.6. This shift will affect the length of the ablation season which will extend to May–September and possibly to the late April–early October. The nature of precipitation will also change. The modelled data show that during the baseline period snow accounted for 85% in April, 59% in May, 76% in September and 97% in October. In 2071–2100, the fraction of solid precipitation is projected to decline to 52–75% in April and October and less than a third of total precipitation in September and May which is characteristic of June and August of the baseline period.

5. Discussion

[27] There is a clear trend for glacier recession in the Russian Altai between 1952 and 2004. The surface area loss of 19.7 ± 5.8% is in agreement with 10–30% glacier surface area reduction between the late 1940s and 2000 in the four glaciated massifs of the Mongolian Altai [*Kadota and Gombo*, 2007]. Three of the Mongolian massifs accommodated mostly valley glaciers, which are typical of the Chuya Ridges, and exhibited the average surface area loss of 17% that is in very close agreement with our results. The fourth, dominated by the flat-summit glaciers that are more sensitive to ELA changes [*Kutuzov and Shahgedanova*, 2009] exhibited higher retreat rates. This is the only other study of the recent glacier change in the Altai, which uses similar data, methods, and glacier sample. Other studies report slower glacier retreat, however, they are based either on smaller samples or on different methods. *Surazakov et al.* [2007] estimated that seven glaciers of the Aktru basin (Figure 1) have lost 7% of their area in 2006–1952 with individual retreat values ranging between 5% and 29%. These glaciers are located at higher than average altitudes which partly explains the lower retreat rates. Our results for the same glaciers based on the 1952 values published in *CG* [1974] indicate a greater retreat of 12%. The difference appears to be within the uncertainty of our estimations. A number of glaciers in this 7-glacier sample have features complicating their mapping including glaciers with very steep slopes (e.g. Karatash and Malyi Aktru) and a debris-covered Pravyi Aktru. Similarly to our results for the 21 test glaciers (Table 2), this comparison points at an overestimation of the 1952 glacier area in the *CG* [1974, 1977] confirming the importance of using large glacier samples to reduce uncertainty due to mapping errors and local factors.

[28] The mass balance and areal extent of the Altai glaciers fluctuate in response to changes in the JJA air temperatures and annual precipitation. Both statistical [*Narozhny et al.*, 2005] and modelling [*De Smedt and Pattyn*, 2003] studies and results of ice-core analysis [*Nakazawa and Fujita*, 2006] identify summer air temperature as the main factor driving the behaviour of the Altai glaciers. The 20th century warming was delayed in the Altai in comparison with regions located further west, e.g. the European Alps [*Diaz and Bradley*, 1997] and the Caucasus [*Shahgedanova et al.*, 2009]. However, a strong positive trend in JJA temperatures was observed in the Altai since the mid-1980s and although it was more pronounced below 2000 m a.s.l., it characterized the high-altitude areas too [Figures 5a, 5c; *Narozhny and Adamenko*, 2000]. Between the 1950s and 1994, there has been no clear trend in annual precipitation in the Russian Altai [*Narozhny and Adamenko*, 2000] and the 1940–2000 times series from the Mongolian Altai also indicate strong interannual variability but no significant long-term trends in precipitation [*Kadota and Gombo*, 2007]. The more recent data for the Russian Altai is limited to two stations only and of these only Kara-Tyurek exhibits a weak positive trend in precipitation. Furthermore, the ice cores from Belukha Glacier in the Katun Ridge of the Russian Altai indicate a strong increase of 1.6 ± 0.4°C in JJA temperatures between 1811 and 2000, which intensified in the 1950s and especially since the 1980s, but no change in accumulation [*Henderson et al.*, 2006]. The observed summer warming appears to be a driving force of glacier retreat in the Altai between the 1950s and 2000s. A brief period in the 1980s when negative anomalies in summer temperatures were observed resulted in a short-term mass gain by the Altai glaciers [*Dyurgerov and Meier*, 2000; Figure 7c]. An additional factor is a decline of frequency of solid precipitation in the warm season accompanying climatic warming. Summer snowfalls limit the overall summer melt and their frequency has declined particularly since the 1980s [*Narozhny and Adamenko*, 2000]. The modelled data show that in 1961–1990, 20% of total precipitation was accounted by snow in July and this will be reduced to less than 2% in 2071–2100.

[29] The JJA mass balance values (mostly ablation) on the Malyi Aktru Glacier were above average since the 1990 (Figure 7a) exceeding the record average by two standard deviations in the warmest on record summer of 1998 and in 2008 (Figures 5a, 5c). The September–May mass balance values (mostly accumulation) (Figure 7b) were above average in 1985–1990 but declined in the 1990s. The cumulative mass balance declined strongly after 1998 (the warmest year on record) while ELA increased from an average of 3130 m a.s.l. in 1961–1987 to 3230 m a.s.l. in 1988–2008 (Figures 7c, 7d). This yields a rise of 100 m per 0.8–0.7°C warming observed at the Akkem and Kara-Tyurek stations or, after interpolation, a rise of 120–140 m per 1°C warming. This is similar to the ELA sensitivity to temperature change in the Swiss Alps estimated as 120–170 m per 1°C warming [*Paul et al.*, 2007]. The average projected JAA temperature increase is 4.5°C and 6.7°C under the B2 and A2 scenarios. Assuming unchanging

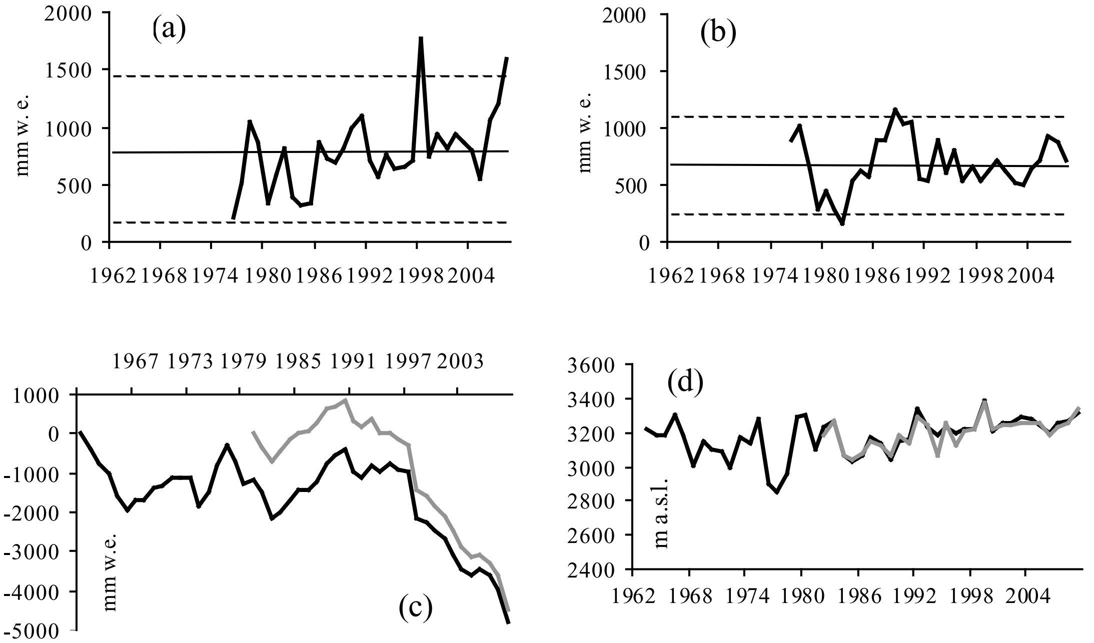

Figure 7. June–August (a) and September–May (b) mass balances for Malyi Aktru Glacier, cumulative mass balance (c), and ELA (d) for the Malyi Aktru (black line) and Levyi Aktru (grey line) glaciers. Solid horizontal lines show record averages; dashed horizontal lines show ± two standard deviations. Mass balance year starts in October and ends in September of the following calendar year. The end (January–September) calendar year is shown on X-axes.

ELA sensitivity, by the end of the 21st century, it should increase by 540–630 m and 800–940 m respectively. An important implication of the ELA increase is a decline in accumulation area ratio (AAR) and a subsequent increase in sensitivity of glaciers to climatic warming [*Dyurgerov*, 2003]. Only net mass balance and ELA records are available for Levyi Aktru and these correlate strongly with the Malyi Aktru series with correlation coefficients of 0.96 and 0.93 respectively (Figures 7c, 7d).

[30] Records of glacier front fluctuations provide further insight into glacier changes in the 20th century (Figure 8). The Malyi Aktru Glacier was retreating at a rate of 3–10 m a^{-1} between 1911 and 1998 with the slowest retreat in the 1940s and late 1960s and more active retreat in the 1950s [*Revyakin and Mukhametov*, 1981; Figure 8]. Between 1998 and 2008 its retreat rates increased to an average rate of 15 m a^{-1} ranging between 12 m a^{-1} and 22 m a^{-1}. A change in glacier front positions of the Malyi Aktru and Sofiyskyi glaciers in reaction to a step change in mass balance occur with a delay of 10–12 and 18–20 years respectively [*De Smedt and Pattyn*, 2003]. Therefore, the currently observed rapid retreat of Malyi Aktru can be seen as a response to the anomalously high temperatures and ablation of the late 1990s (Figures 5 and 7). A rapid retreat of the Sofiyskyi Glacier occurred in 1952–1963 at a rate of 21 m a^{-1} slowing down to 7 m a^{-1} in 1963–1997 but intensifying to 15 m a^{-1} in 1997–2000 [*De Smedt and Pattyn*, 2003]. *Surazakov et al.* [2007] reported an acceleration of the retreat of the glaciers in the Aktru valley after 1975 in comparison with 1952–1975. In this respect, the behavior of glaciers in the Chuya Ridges is different from the Mongolian Altai where glaciers retreated actively either in the 1950s or in the 1970–1980s but remained stationary in the 1990s [*Kadota and Gombo*, 2007].

[31] The regional climate scenarios show an annual precipitation will increase by 15% and 5% for A2 and B2 scenarios in 2071–2100, however, partly this increase will be offset by the increasing proportion of liquid precipitation in the main accumulation months of April, May and October. A 25–40% increase in precipitation is required to offset the effects of a 1°C warming on glacier mass balance and subsequent retreat [*Braithwaite et al.*, 2002; *Oerlemans*,

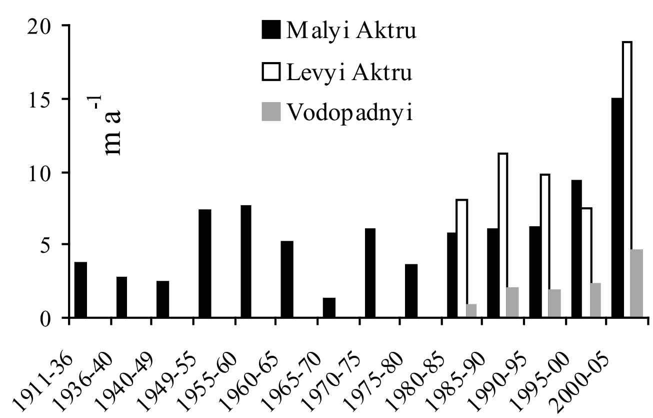

Figure 8. Average retreat rates of glacier fronts. The post-1980 data are from Fluctuations of Glaciers [*WGMS*, 2008 and earlier issues]. The pre-1980 data for Malyi Aktru are from *Revyakin and Mukhametov* [1981]. Note that the pre-1949 data are averaged over periods of different length.

2005]. The projected precipitation change will not compensate for the projected JJA warming of 6–7°C and 3–5°C and increase in the length of ablation season implying further glacier retreat. A modeling study of dynamic sensitivity of Sofiyskyi Glacier to summer warming by *De Smedt and Pattyn* [2003] indicated that this glacier, currently one of the largest in the Altai, will vanish by the end of the 21st century should summer air temperatures rise by more than 5°C. This is a JJA warming simulated for the most aggressive A2 group of scenarios. The Sofiyskyi Glacier, however, exhibits higher sensitivity to temperature changes than some smaller glaciers and future shrinkage of other glaciers may be slower.

6. Conclusions

[32] In this study, glacier surface area in the North and South Chuya Ridges of the Altay Mountains was evaluated using 2004 ASTER imagery and glacier retreat since 1952 was assessed using data published in the *Catalogue of Glaciers of the USSR* [1974, 1977]. The glaciated area has declined and a number of glaciers have fragmented due to the increasing summer temperatures. 126 glaciers (not smaller than 0.5 km^2 in 1952) have lost 19.7 ± 5.8% of their net surface area. Providing that an increase in summer temperatures started in the Altai after the 1980s and glacier front fluctuations and mass balance records for individual glaciers show that their shrinkage accelerated since the 1990s, it can be concluded that glaciers of the Altai reacted rapidly to the observed climatic warming. The projected strong increase in summer temperatures and only weak increase in precipitation accompanied by an increase in the fraction of liquid precipitation in late spring and early autumn are likely to result in their further retreat in the 21st century. Modeling future changes in the extent of glaciated area in the Altai using the output from PRECIS represents direction for future research.

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