



Measurement and Assessment of Snowmelt Erosion in Western Siberia

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Abstract

Snowmelt erosion is the most destructive type of soil degradation in the dissected territories of Western Siberia. More than 50% of all Siberian farmlands are subject to erosion to various degrees. Issues related to the processes and consequences of soil erosion are a serious problem for both scientists and farmers. The purpose of this work is to present the main methods and devices used to define the quantity and quality of snowmelt water surface runoff and the damage caused by the snowmelt soil erosion. In Western Siberia, one of the main factors of erosion is snow cover, which forms and accumulates for up to five consecutive months. The water content in snow ranges from 65 mm in very low-snow winters to 255 mm in extremely high-snow winters. To measure soil erosion during the snowmelt, field methods were used to determine parameters such as the meltwater runoff volume, number of days with active meltwater drainage, water flow temperature and turbidity

of the snowmelt waters. Diagnostic signs of the effects of soil erosion after snowmelt included the volume of the soil solid phase removed by meltwater and the change in thickness of the humus horizon. The losses from erosion after winters with different amounts of snow were: solid phase of the soil—from 1 to 11 t/ha⁻¹, carbon—from 5 to 1000 kg⁻¹, and nitrogen—from 1 to 55 kg⁻¹. The described methods for studying soil erosion during the snowmelt are simple to use and do not require expensive equipment in the field.

Keywords

Soil erosion · Snowmelt surface runoff · Slope gradient · Western Siberia · Soil loss · Humus and nutrient removal · Chernozem transformation

7.1 Introduction

The land degradation process due to the rate of soil loss exceeding the soil formation contributed to the formation of the modern physical landscape (Alewell et al. 2015). Soil erosion is among the eight soil threats listed in the Soil Thematic Strategy of the European Commission (Panagos et al. 2015). Erosion is the most common and dangerous type of soil degradation in the world. Soil degradation is inherently a gradual process, and the consequences are not always

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obvious until damage is done. Erosion processes have disastrous consequences for humans and the environment, since they threaten the existence of soil as the main means of agricultural production and an indispensable component of the biosphere (Dobrovolskii 1997). The current rates of soil degradation continue to exceed the rates of soil formation, and soil degradation is jeopardising basic food production capabilities in certain parts of the world, even in the short term (Scherr 1999). Globally, 80% of agricultural land is subject to moderate and severe erosion, with soil loss of about 30 t ha⁻¹ per year on average (Pimentel 2006). Soil erosion reduces agricultural productivity and is estimated to reduce global gross annual agricultural output by 10% (Lal 1998).

Soil erosion is clearly affected by the amount and intensity of precipitation (Li and Fang 2016; Shi et al. 2020). Particularly, in mid-high latitudes and cold regions such as the northern parts of North America, Europe, Russia and northeast China, soil erosion depends predominantly on the climate (Edwards et al. 1998) and vegetation characteristics (Maltsev and Yermolaev 2019). Seasonal variations characterise the degree of snowmelt soil erosion due to the increasing surface runoff caused by melting snow in early spring and intense rainfall in summer (Lal 1998; Tanasienko 2003; Tanasienko et al. 2009; Ouyang et al. 2017; Wu et al. 2018). Snowmelt-induced gully erosion is one of the most harmful forms of degradation, particularly on arable lands (Maltsev and Yermolaev 2020; Xu et al. 2019). In addition, repeated freeze–thaw cycles significantly contribute to gully erosion (Luffman and Nandi 2019). Different factors contribute to meltwater erosion, including the slope gradient and flow rate (Shi et al. 2020); the topography and microtopography (Barneveld et al. 2019); climatic factors (temperature and the amount, rate and repeatability of snow melting) (Luffman and Nandi 2019; Tanasienko et al. 2019; Xu et al. 2019); the type of vegetation (Maltsev and Yermolaev 2020); and the depth of frost penetration, slaking and other aspects (Barneveld et al. 2019).

In Western Siberia (Russia), Chernozem soils (13.3 million ha) occupy the southern part, over a relatively flat territory that is completely cultivated with crops (10.7 million ha, Khmelev, Tanasienko et al. 2009). The further expansion of arable land is possible only on soils located on slopes. Slopes with gradients exceeding 3° are more susceptible to erosion than flat lands. Approximately 30% of these sloped lands are located on the northern slopes, while 50% are located on the southern slopes. On average, 10% of all arable land is located on slopes with a gradient of 6–9°, and approximately 5% is located on slopes with a gradient exceeding 9°. Soils on such slopes are exposed to very strong sheet erosion. Predsairye, Priobye and the Kuznetsk Depression are areas in the southeast of Western Siberia with almost 2.5 million hectares of arable land. More than 60% of these soils are Chernozems. About 20% of the arable land of this region has already been eroded to varying degrees. Therefore, the study of erosion processes and the causes of their various developments in this area appear to be highly relevant.

Despite advances in understanding rain erosion, the issues of snowmelt erosion remain poorly investigated. In particular, the relationships between rainfall, snowmelt water and water discharge on the deep and long-term frozen soils of Western Siberia are not fully understood. The mechanisms governing the detachment and sedimentation of soil particles are also poorly understood. There is also scant information on field observations in the zones of snowmelt erosion. Thus, the specific objectives of this study were to identify and assess the causes of the development of erosion processes and their impact on soils of Chernozems in years with different amounts of snow in the southeast of Western Siberia.

7.2 Study Regions and Relief Characteristics

The study area is located in Western Siberia and is a zone containing extremely fertile Chernozem soils and intensive agriculture. The regions

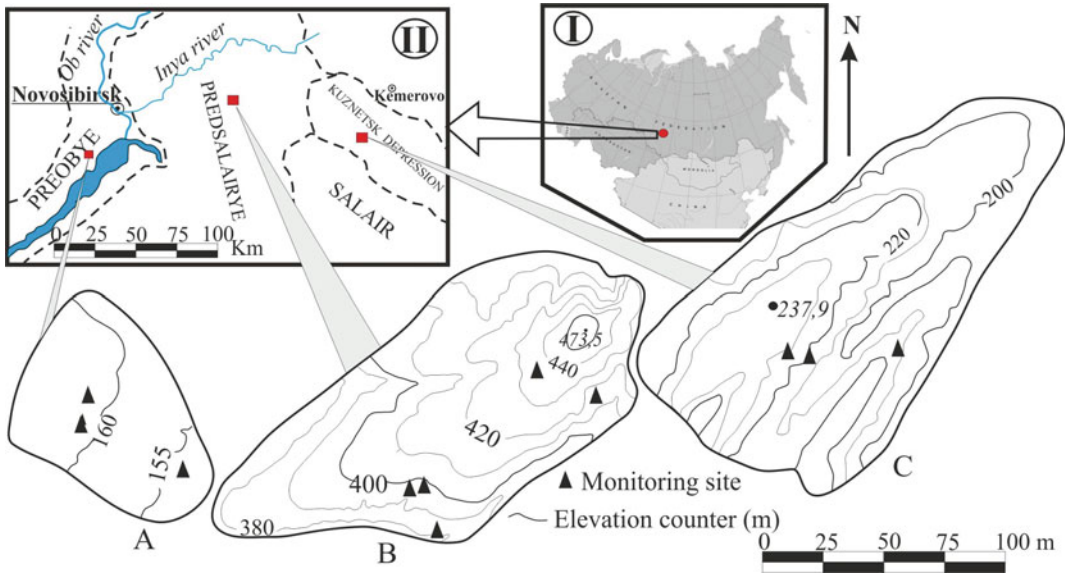


Fig. 7.1 Map of monitoring sites at Priobye (a), Predsairye (b) and the Kuznetsk Depression (c). The insert shows the location of the study area in Russia (I), southern Western Siberia (II) (Tanasienko et al. 2011)

studied include Novosibirsk (Predsairye and Priobye) and the Kuznetsk Depression (Kemerovo region) (Fig. 7.1). The areas differ in terms of their geomorphological and climatic characteristics and soils.

Predsairye is a flat, partly hilly, highly incised plain, located in a forest steppe zone. Arable land occupies more than 60% of all agricultural land, or approximately 800,000 ha; 23% of this territory is already eroded to a slight to moderate degree. The very high horizontal and vertical incision of the territory promotes erosive processes (Barneveld et al. 2019). Within Predsairye, subareas with varying degrees of erosion are easily distinguishable. The most severe erosion processes in Predsairye were observed in the Bugotac Hills subarea, which is a series of small hills extending in a northeastern direction.

The hill height is 480 m above sea level. The horizontal incision of this territory is the greatest for Predsairye and ranges from 1.5 to 2.2 km km⁻¹; the vertical incision is 75–100 m. Due to the strong incision of the territory, the basic landscape elements are slopes with gradients of 9–12° near the watershed area and 25–30° closer to gullies. The convex slopes are predominant

forms in this region; therefore, the greatest intensity of water discharge mainly occurs in the middle and bottom portions of the long slopes. Slopes with a southerly aspect are the steepest, while slopes with a northerly aspect are fairly steep and therefore less subject to erosion processes. The Bugotac Hills are classified as a territory with a very high risk of erosion.

The Kuznetsk Depression is also located in the forest steppe zone, with a strongly incised relief. The horizontal incision varies from 0.6 to 0.8 km km⁻¹ in the west of the depression and from 1.0 to 2.6 km km⁻¹ in the rest of the territory, while in some agricultural areas, the horizontal incision reaches 3.3–3.5 km km⁻¹. The vertical incision of the territory is similar to Predsairye (75–100 m). Due to such incisions, the watersheds occupy approximately 20% of the territory. In the incised parts of the depression, arable land occupies slightly more than 1 million ha, 15% of which is eroded from a slight to a moderate degree.

Priobye is located in a transition from a forest steppe to a steppe zone with absolute heights of 130–310 m, good natural drainage, an absence of primary salinity and deep ground waters (10–

15 m). Priobye's relief has a slight horizontal incision of 0.6–0.8 km km⁻¹. The watersheds are flat and occupy nearly 40% of the territory. The slope steepness in the watersheds ranges from 15° to 30° and that near gullies reaches 25–30°, indicating a significant erosion hazard in the territory. Of 180,000 ha of arable land, 15% are already eroded to a slight or moderate degree. Thus, this territory was classified as at a moderate risk of erosion (Nikitenko 1963).

7.3 Key Plot Characteristics

The theory of studying erosion is based on general models of the development of the biosphere and landscape elements. Studies of erosion processes should be based on a systematic approach using the principle of landscape catenae, and the method of field observations is the main tool for their study. To study the development of soil erosion in Western Siberia, it is advisable to place key plots in various landscape positions: near watersheds (non-eroded or virgin soils); in transluvial positions (weakly, moderately and strongly eroded soils) and in transaccumulative positions (accumulated soils).

Our studies of the factors and the process of erosion were carried out mainly on the slopes with a southeastern exposure, since these slopes are most susceptible to erosion. According to the principles of landscape catenae, non-eroded soils occupy a flat watershed, below which there are relatively long (600–800 m) dissected slopes. The upper boundary of weakly eroded soils ran approximately 150 m from the boundary with the watershed, and the lower one reached 300–350 m from this boundary. Areas with a slope of 3–6° were occupied by moderately eroded soils. Below the slope, strongly eroded soils were found with an upper boundary at 450–500 m from the watershed boundary. The southeastern slope ranged from 7 to 9°.

On the slopes, along with erosion processes, accumulative processes also took place. At the bottom of the erosion furrow, from the moment of its formation, a partial accumulation begins (Makaveev 1955). In this case, there are three

zones of sediment accumulation. The first is arable land, where there is intense deposition of sediments in places where the slopes go down below the confluence of temporary streams, in front of natural and artificial barriers. Golosov (1988) made the assumption that on average, up to 20% of the soil carried by the thawed soil from the slopes settles within the watersheds.

The second zone of accumulation is below the border of agricultural lands. Here, the soil material is re-deposited due to the dispersion of the water flow at the field exit and due to the increased roughness of the underlying surface. Fan formation occurs due to its expanding and removing the outer edge from the arable land. In this zone, depending on the rate of flushing, 10–75% of sediment is retained. First of all, particles drawn in by the flow accumulate, the proportion of which increases with an increase in the genuine roughness of the underlying surface. The remaining sediments are carried by the stream to ground-covered areas with slopes of 30 m or more from the borders of agricultural fields (third zone). In this zone, the depth of sediments depends on the turbidity of meltwater. Thus, in the process of erosion, the soil-removing material partially accumulates in the lower parts of the slopes, where mainly semi-hydromorphic soils are formed. In this part, in most cases, the humus horizons are 50% thicker than in upland soils. To obtain a representative material, it was necessary to select a highly dissected area of 1–1.5 km² followed by a geographical survey using both a comparative–geographical method and an analysis of the soil cover structure to determine the soil geochemical characteristics within the landscape.

7.4 Climatic Conditions of Study Sites as Related to Snowmelt Erosion

Precipitation in Western Siberia in the warm period is 200–350 mm, but only heavy rain has an erosive effect. However, since precipitation occurs in Western Siberia in a narrow time interval, on an annualised basis, summer

precipitation causes short-term soil erosion. Due to the partial protection of the soil by plants, the erosion effect from heavy rains is much weaker than from the snowmelt water in spring. In Western Siberia, soil erosion develops mainly under the influence of meltwaters, since the cold period lasts up to 6 months, and fallen snow with a depth of 70–200 mm melts in a very short time (on average 5–7 days). The development of soil erosion in this area depends on (1) the air temperature in both cold and snow melting periods; (2) the amount of solid precipitation in the cold period of the year (November–March); (3) soil moisture before winter and immediately before snow melting; (4) the depth of soil freezing; and (5) weather conditions of snowmelts.

7.4.1 Air Temperature in Pre-winter, Winter and Early Spring

One of the main factors that determine the depth and duration of soil freezing is the air temperature before the winter, during the winter and in early spring. In Western Siberia, the pre-winter period is the period from the moment the average daily air temperature passes 0 °C until the establishment of a stable snow cover. Winter (cold period) is the period from November to March, when the average daily air temperature is below 0 °C, and a permanent snow cover forms on the soil surface. Early spring is the period from the last 10 days of March to the second 10 days of April, when the daytime air temperature rises above 0 °C, and the night-time temperature remains below zero. By the end of early spring, the average daily air temperatures rise above freezing point.

In the study area, the air temperature in the first half of October is predominantly above freezing. However, in other years, average daily air temperatures sometimes remain stably below 0 °C before the formation of permanent snow cover. In this case, the upper 30 cm layer of arable soil quickly freezes, and the deep snow formed by mid-December is no longer able to prevent the subzero temperatures from penetrating deep into the soil. During such cold periods,

arable soils can freeze deeper than 160 cm. The air temperature remains significantly below zero without thaws throughout the entire cold period (November–March).

The coldest month is January, but in spring there are frequent frosts, since the spring rise in air temperature is often interrupted by a sudden (but short) sharp cold snap. Thus, the air temperature is observed to be relatively low (but above 0 °C) in April in very snowy winters in Predsairye and the Kuznetsk Depression (+0.8 and +0.7 °C, respectively).

As shown in Table 7.1, both in the cold period and in early spring, the lowest air temperature was recorded in the Kuznetsk Depression. This can be explained by the fact that this geomorphological region is located in the east of Western Siberia. Also, the end of the cold period (March) in this area is always characterised by an air temperature well below zero, from –7.1 to –11.7 °C, despite the snowy winter. In addition, the cold period in the Kuznetsk Depression is about 10–15 days longer than in Predsairye and Priobye.

The air temperature in individual months with different periods of snowiness varies greatly (Fig. 7.2). The measured data showed that the Kuznetsk Depression was the coldest site; there, the sum of subzero air temperatures ranged from –2115 to –2660 °C in each hydrological year. Winter in Predsairye was warmer than in the Kuznetsk Depression. These data may indicate deep and severe freezing of the soil in winters with very shallow snow and very snowy winters. This is one of the reasons for the formation of meltwater surface runoff in spring.

Thus, the less snow falls in November and, in general during the cold period of the hydrological year, and the lower the subzero air temperatures on these dates, the deeper the arable soil freezes. Conversely, a deeper snow cover and relatively low frequency of subzero air temperatures in the pre-winter period lead to a warmer soil profile. The depth of freezing of forest steppe soils ranges from 220 cm in very low-snow winters to 90 cm in very snowy winters, i.e. depending on the snowiness of the hydrological

Table 7.1 Average long-term air temperature in different hydrological years in the pre-winter, winter and early spring periods

Weather station (WIGOS Station Identifier ^a) Data for years	Characteristics of hydrological years	n	Average monthly air temperature (°C)											
			Months											
			X	XI	XII	I	II	III	IV					
<i>Predslaitrye</i>														
TOGUCHIN (0-20.000-0-29,636) 1960–2018	Very low-snow	4	3.7 ± 0.9	-8.0 ± 1.2	-14.6 ± 1.4	-19.2 ± 0.8	-16.2 ± 1.1	-7.1 ± 2.1					3.6 ± 0.7	
	Low-snow	10	1.1 ± 0.8	-6.7 ± 1.0	-13.7 ± 1.1	-14.1 ± 1.3	-16.0 ± 1.3	-9.0 ± 0.9					1.4 ± 0.7	
	Normal-snow	19	1.3 ± 0.6	-8.7 ± 0.8	-14.3 ± 0.9	-17.5 ± 0.8	-14.9 ± 0.9	-8.3 ± 0.7					1.3 ± 0.6	
	High-snow	17	2.5 ± 0.3	-7.0 ± 0.9	-13.8 ± 1.1	-17.1 ± 1.4	-15.4 ± 1.1	-6.9 ± 0.7					1.9 ± 0.9	
	Very high-snow	8	0	-6.2 ± 1.4	-15.3 ± 2.9	-18.0 ± 2.8	-15.7 ± 2.3	-9.0 ± 1.2					0.8 ± 0.6	
<i>Priobyie</i>														
ORDYNSKOE (0-20.000-0-29,726) 1956–1988; 2006–2017	Very low-snow	5	-	-4.4 ± 1.3	-14.8 ± 1.4	-14.2 ± 0.8	-17.3 ± 1.5	-8.9 ± 2.3					2.3 ± 0.3	
	Low-snow	6	-	-6.6 ± 1.2	-16.4 ± 1.5	-15.0 ± 1.3	-16.5 ± 1.3	-8.9 ± 1.2					3.6 ± 1.0	
	Normal-snow	15	-	-7.9 ± 1.1	-15.1 ± 1.3	-18.3 ± 1.4	-19.2 ± 0.8	-8.2 ± 0.8					2.6 ± 0.7	
	High-snow	12	-	-7.0 ± 1.3	-11.8 ± 1.2	-18.6 ± 1.4	-14.0 ± 1.0	-8.2 ± 0.8					5.2 ± 0.8	
	Very high-snow	6	-	-5.0 ± 1.1	-14.0 ± 2.6	-18.3 ± 2.2	-16.0 ± 1.7	-9.3 ± 1.3					2.2 ± 1.5	
<i>Kuznetsk depression</i>														
KEMEROVO (0-20.000-0-29,642) 1951–1990	Very low-snow	6	-	-10.3 ± 2.7	-14.2 ± 0.7	-16.1 ± 1.7	-16.1 ± 2.5	-6.0 ± 1.2					-0.1 ± 1.1	
	Low-snow	12	-	-9.5 ± 1.2	-14.9 ± 1.2	-16.4 ± 0.9	-16.0 ± 0.9	-10.4 ± 0.5					0.5 ± 0.8	
	Normal-snow	8	-	-12.1 ± 1.9	-19.3 ± 2.1	-18.6 ± 0.7	-18.9 ± 0.9	-10.1 ± 1.1					1.7 ± 0.7	
	High-snow	8	-	-7.1 ± 1.6	-16.1 ± 1.5	-19.9 ± 1.5	-16.5 ± 1.1	-8.7 ± 0.7					1.3 ± 1.1	
	Very high-snow	4	-	-12.3 ± 1.3	-14.1 ± 3.0	-22.2 ± 1.0	-14.4 ± 2.3	-7.5 ± 1.5					-0.8 ± 1.6	

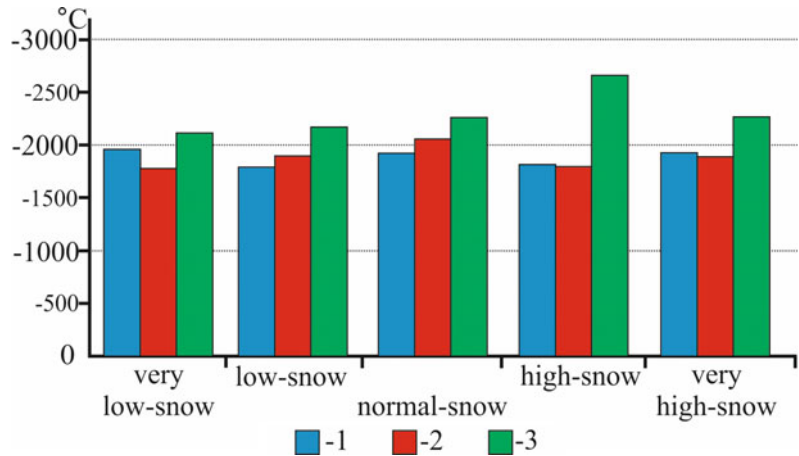
Here and in the other tables: *n* is a number of measurements (years), *Range* is the minimum to maximum values, referring to the whole data pool,

M is an arithmetic mean, *m* is a standard deviation, σ is a mean-square deviation and *V* is a coefficient of variation.

Data for 1956–1960 were from the Reference Book on the Climate of the USSR (1977), those for 1961–1990 from Meteorological Monthly (1966–1990), 1991–2018 data were obtained by the authors

^aWIGOS Station Identifier is the weather station's number in the World Meteorological Organization's Integrated Global Observing System (OSCAR/Surface 2019)

Fig. 7.2 Distribution of subzero air temperatures during the cold period (November–March) in different hydrological years ($^{\circ}\text{C}$), 1—Predsalsairye; 2—Priobye; 3—Kuznetsk Depression



year, the freezing depth of Chernozem soils can more than double.

The average annual data on air temperature in the Kuznetsk Depression, Predsalsairye and Priobye were calculated using data from Internet sources (Weather archive. <https://rp5.ru>, <http://aisori.meteo.ru/ClimateR>) and specialised reference books (USSR Climate Handbook 1977). Since 2005, air temperatures have been recorded at key points in Predsalsairye using autonomous temperature recorders (Thermochron iButton Device 2019). A wooden pole was installed on the thermometric plot, on which a temperature recorder was installed at a height of 2 m from the soil surface. The air temperature recorder was protected from direct sunlight. The temperature recorders were programmed to measure the temperature at regular intervals. In total, over the cold period of a hydrological year and a period of snowmelt (November–April), more than 1000 data points on temperature were obtained.

7.4.2 Precipitation During the Winter

Solid precipitation in Predsalsairye, Priobye and the Kuznetsk Depression falls unevenly during the cold period (Table 7.2). In these regions, regardless of the winter snowfall, the greatest amount of precipitation falls in November, and the least in February. Based on the changes in the amount of solid precipitation during the cold

period (50–220 mm), we identified six types of snowiness in the cold period of the hydrological year. If less than 65 mm of solid precipitation falls during the cold period, the year is classified as a very low-snow year. If the amount of precipitation during the cold period ranges from 65 to 95 mm, the year is classified as a low-snow year. A year with 95–125 mm of solid precipitation during the cold period is referred to as a normal-snow year. A year with 125–155 mm is considered to be a high-snow year. In the study area, during the last 15 years, cold periods prevailed with solid precipitation from 155 to 185 mm, i.e. very high-snow years. During our research, we observed winter periods with snow accumulation of more than 185 mm. We have characterised such cold periods as extremely high-snow years.

According to our research, as well as the Reference Book on the Climate of the USSR (1977), November in very low-snow and low-snow years in any region of Western Siberia was characterised by a fairly low amount of solid precipitation below 16–25 mm. In normal and high-snow years, the snow water equivalent (SWE) increased to 28–30 mm, and in years with very high snow, it rose to 42–60 mm. Therefore, at the beginning of the cold period, in very low-snow and low-snow years, the amount of solid precipitation was half that in very high-snow years.

However, given that in November in very low-snow years, the sum of subzero air

Table 7.2 Precipitation in the pre-winter, winter and early spring periods in different hydrological years

Characteristics of hydrological year		Average monthly precipitation, mm											
n	X	XI	XII	I	II	III	IV						
<i>Predsalairye</i>													
Toguchin weather station (1938–2017)													
Very low-snow	4	50.9 ± 7.5	19.3 ± 3.1	9.4 ± 1.8	11.0 ± 2.9	6.8 ± 2.2	15.4 ± 4.3	20.6 ± 6.3					
Low-snow	23	39.3 ± 3.4	22.8 ± 1.5	22.0 ± 2.1	17.3 ± 1.5	8.9 ± 1.0	10.0 ± 1.0	26.6 ± 3.0					
Normal-snow	24	36.7 ± 3.2	34.1 ± 2.3	28.7 ± 3.1	16.9 ± 1.8	15.1 ± 1.5	14.7 ± 1.5	23.7 ± 2.6					
High-snow	17	46.1 ± 2.7	47.2 ± 3.5	36.5 ± 4.1	16.8 ± 2.1	16.1 ± 3.0	21.6 ± 2.6	26.5 ± 3.1					
Very high-snow	7	48.0 ± 8.7	46.0 ± 6.3	39.1 ± 3.9	33.7 ± 6.0	23.0 ± 2.4	22.9 ± 4.7	25.7 ± 4.1					
Extremely-high-snow	4	59.5 ± 12.2	48.3 ± 3.8	46.2 ± 5.5	31.8 ± 8.1	32.2 ± 2.1	32.8 ± 2.5	18.7 ± 0.5					
<i>Priobye</i>													
Ordynskoe weather station (1936–1988; 2006–2017)													
Very low-snow	8	24.1 ± 2.5	17.9 ± 2.1	13.7 ± 2.6	8.2 ± 0.3	7.1 ± 1.4	11.5 ± 3.3	20.0 ± 4.3					
Low-snow	10	34.6 ± 5.5	21.2 ± 1.1	23.1 ± 2.2	13.2 ± 2.0	5.3 ± 1.1	11.2 ± 1.2	18.2 ± 4.8					
Normal-snow	26	37.9 ± 3.0	26.6 ± 2.2	26.1 ± 2.3	20.8 ± 1.8	17.1 ± 1.3	14.9 ± 1.8	25.5 ± 2.7					
High-snow	11	41.7 ± 7.3	47.9 ± 6.0	27.8 ± 4.0	27.8 ± 4.0	20.3 ± 3.4	18.8 ± 2.8	23.7 ± 3.8					
Very high-snow	4	40.3 ± 3.7	50.0 ± 7.2	33.3 ± 9.2	37.5 ± 5.4	25.0 ± 5.4	22.7 ± 4.2	23.8 ± 6.4					
<i>Kuznetsk depression</i>													
Kemerovo weather station (1938–2017)													
Very low-snow	19	25.7 ± 3.3	27.6 ± 3.2	12.7 ± 1.5	9.5 ± 1.0	6.8 ± 0.9	7.0 ± 1.0	17.0 ± 1.8					
Low-snow	17	25.9 ± 3.9	22.2 ± 1.7	19.3 ± 1.7	17.1 ± 2.2	10.4 ± 1.4	8.8 ± 1.2	19.5 ± 4.6					
Normal-snow	8	37.2 ± 4.4	33.8 ± 3.2	24.1 ± 4.3	24.1 ± 4.5	15.0 ± 3.0	12.9 ± 3.0	16.1 ± 2.7					
High-snow	14	10.0 ± 11.5	42.4 ± 3.9	33.0 ± 2.6	25.4 ± 3.2	16.7 ± 2.6	18.2 ± 2.2	23.0 ± 5.9					
Very high-snow	13	–	48.6 ± 3.1	45.9 ± 3.6	27.0 ± 3.2	23.0 ± 2.8	22.6 ± 3.6	–					

temperatures in Predsairyie and the Kuznetsk Depression ranged from -240 to -295 °C, in practice the shallow snow cover did not protect the soil from subzero temperatures. Within the forest steppe zone of the southeast of Western Siberia, the period during which permanent snow cover is established has a great influence on the formation of snowmelt water surface runoff and, as a consequence, on the water regime of eroded Chernozems. The formation of the first temporary snow cover occurs at the beginning of the pre-winter period. This period often coincides with a steady air temperature transition through 0 °C towards subzero temperatures. It should be mentioned that the prolonged pre-winter phase leads to the unhindered penetration of subzero temperatures into the soil profile. According to 'Meteorological Monthly' (1966–1990), snow most often falls on frozen soils in Priobye (71%), and in Predsairyie and the Kuznetsk Depression (62%).

The formation of permanent snow cover is determined by the conditions of the underlying surface. If snow falls on thawed soil with a surface temperature varying from 3 to 5 °C, the snow melts relatively quickly and replenishes moisture in the upper part of the humus horizon. In this case, the gap between the appearance of the first snow and the formation of a stable snow cover is quite significant. If the first snow falls on the surface of frozen soil, then the formation of stable snow cover is controlled only by the weather conditions at that time of the year. Therefore, in the pre-winter period, when determining the moisture in the profile of non-eroded and eroded soils, one should take into account factors such as the beginning of the formation of stable snow cover.

The distribution of snow also depends on the frequency of snowstorms, which contributes to a considerable transfer and a greater compaction of snow. Freshly fallen snow is usually blown off flat and convex relief elements and accumulates in depressions and by ridges. Snowfall reaches its maximum values in Western Siberia at the end of February and beginning of March, when wind activity increases markedly. Our studies of the distribution of snow cover were performed on

slopes with southeast exposures as these slopes are most susceptible to erosion processes. Since Predsairyie and the Kuznetsk Depression are characterised by a marked horizontal and vertical division of the territory, the distribution of snow here plays an important role in the development of soil erosion. On the erosive catena, the maximum thickness of the snow cover is inherent in the eluvial position of the landscape (watershed space). This part of the catena is always dominated by non-eroded soils, which in our study are represented by virgin and arable lands.

The presence or absence of vegetation also strongly affects the distribution of snow cover. Thus, the virgin land located on the slope with southern exposure is characterised by a greater snow cover than arable land occupying an eluvial landscape position. For example, in the Predsairyie region during the high-snow period in 2008/2009 and in the extremely high-snow years of 2014/2015, the depth of snow cover in the virgin lands was 20% higher than in the catchment area occupied by arable non-eroded soil. The southern arable slopes, regardless of the snowiness of the hydrological year, always accumulate less snow than flattened non-eroded areas and especially slopes with a cold orientation (Rikhter 1945).

Studying the snowmelt water runoff requires accurate information about the snow water equivalent (SWE), which is the total amount of water in solid and liquid forms before the spring flood. The distribution of snow cover and SWE were determined by continuous snow surveys before the snow melted (in the south of Western Siberia, the second and third 10-day periods in March) along parallel routes crossing the catchment every 100 m. The height of the snow cover was determined with a metal snow scale (160 cm long) every 5 m. The density of the snow cover (g cm^{-3}) and the water equivalent of snow (mm) were determined using a BC-1 snow gauge every 100 m in duplicate (Fig. 7.3). The snow gauge records the snow water equivalent at a specific observation point, as well as the height of the snow cover.

Using data on snow water and snow depth, the snow density was calculated. The SWE in the

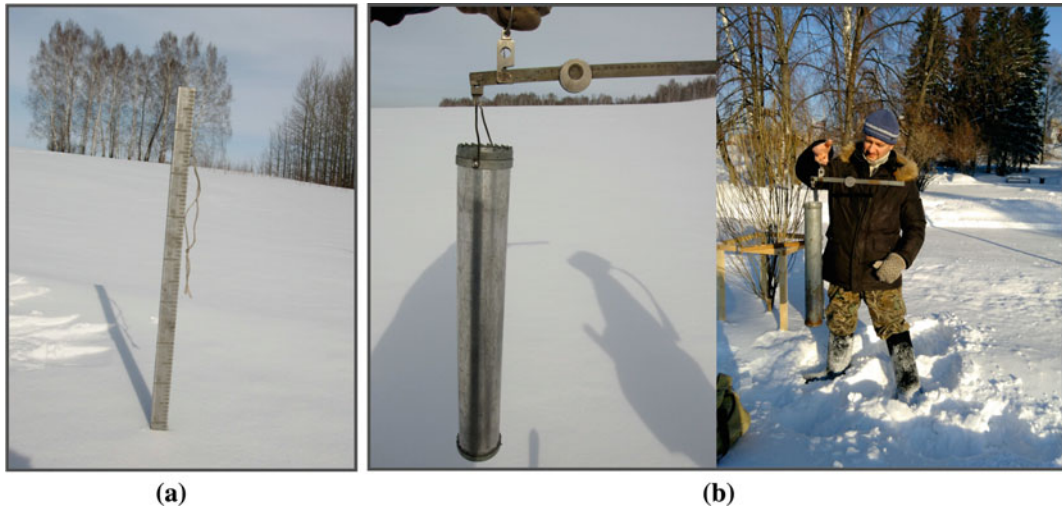


Fig. 7.3 Determining the snow depth with a metal snow scale (a) and calculating the snow water equivalent (SWE) by BC-1 (b) during the snow survey stage

catchment area was determined by multiplying the arithmetic mean of the snow density by the average snow depth on a particular element of the catchment area (Chumbaev and Tanasienko 2016).

7.4.3 Soil Moisture in the Pre-winter Period, During the Melting of Snow

To detect whether vaporous moisture was frozen in the lower part of the soil profile during the cold period of the hydrological year, and to register the redistribution of moisture along the soil profile, it was necessary to determine the amount of moisture in the profiles of non-eroded and eroded soils in the autumn–winter period. Volumetric soil

Fig. 7.4 Soil sampling for water content



samples were taken at three positions on each plot at different depths in steps from 10 cm to 1.5 m (Fig. 7.4). The moisture content of soil samples was analysed in the laboratory by drying at 105 °C to a constant weight (Vadyunina and Korzhagina 1961). Control measurements of soil moisture were carried out immediately before the snow began to melt. After the meltwater surface runoff appeared at the control points, the soil moisture was measured twice a day (morning and evening). When the snow had finished melting, the amount of moisture entering the soil profile with surface runoff was calculated.

The total soil moisture consumption after harvest had decreased. This was due, on the one hand, to the lack of moisture transpiration by cultivated plants, and on the other, to autumn air temperatures not exceeding 10–15 °C. It is desirable for precipitation of 70–90 mm to be retained in September–October in the Chernozem profile at eluvial and transluvial positions of the landscape. The low intensity of precipitation enables it to be fully absorbed by the soil (Tanasienko et al. 2019). The maximum moisture content at a depth of 1.5 m (on average 450 mm) in the pre-winter was found in non-eroded Chernozems (virgin and arable soils) in Predsairyie (Table 7.3). In the eroded soils, it was 104–129 mm less. The largest pre-winter moisture content was found in the slightly drifting soils located on the slope plume. The Priobye site is a transition from a forest steppe to steppe zone. Consequently, the atmospheric humidity during the cold period of the hydrological year should be much less than in the Predsairyie region. In Priobye, a large moisture deficit was observed both on the eroded and non-eroded arable land in the pre-winter period, ranging from 44 to 62 mm over the years of the study. The water content in the snow in Priobye has fluctuated sharply over the past 80 years (from 57 mm in low-snow years to 181 mm in very high-snow years). Comparison with the autumn moisture content leads us to conclude that under favourable conditions during the snow melting period, practically all meltwater can be absorbed by the profile of arable Chernozems. Non-eroded and eroded

Chernozems of the Kuznetsk Depression showed the lowest amount of autumn soil moisture. By the time a permanent snow cover is created in these soils, a significant moisture deficit is formed, which ranges from 50 to 100 mm.

The soils in Western Siberia remain frozen when the snow starts to melt. It is only when the snow cover is destroyed and thawing patches form that they begin to thaw during the day. A gradual increase in the daytime air temperature and the duration of sunshine leads to deep thawing and the saturation of soil with the meltwater. Complete thawing of the soil in the study region was observed 7–15 days after the disappearance of snow cover. In the soils of Western Siberia, despite the high-snow reserves, after snow melting, there can be both an increase in moisture and a deficit in the soils. We calculated the moisture reserves as the moisture content in the soil layers before and after snowmelt. As can be seen from Table 7.4, after the snow melts in any region of southeast Western Siberia, most of the moisture accumulates in the upper 50-cm soil layer. Total moisture in the upper 50 cm soil layer of virgin lands increases by an average of 20 mm in comparison with autumn. However, in the deep layers of virgin soils in Priobye and the Kuznetsk Depression, the spring total moisture is practically equal to the autumn. In non-eroded arable Chernozems, the accumulation of moisture was recorded in all layers. The non-eroded Chernozems in the Kuznetsk Depression absorbed twice as much meltwater as the Chernozems of Priobye. The relatively small absorption of meltwater by the Chernozems of Priobye is explained by the small amount of snow, the moderately loamy soil texture and the low humus content, which contributed to the soil's good water-holding capacity.

In our study, with an increase in winter snowfall in a 1.5 m layer of weakly eroded Chernozems in Predsairyie, Priobye, and the Kuznetsk Depression, spring moisture increased from 290 to 370 mm. The reason for this wide data distribution is the different accumulation of moisture in the soil in the pre-winter period. The high initial moisture content in the soil profile in

Table 7.3 Pre-winter moisture reserves^a and field water capacity (FWC) in the profile of Chernozems in normal-snow hydrological years in southeast Western Siberia

Degree of soil erosion, type of land use	Layer (cm)	Statistical parameters			FWC (mm)
		<i>n</i>	Range (mm)	$M \pm m$ (mm)	
<i>Predsalairye (1999–2012)</i>					
Non-eroded, virgin land	0–50	10	149–236	187 ± 9.2	157
	50–100	10	108–178	137 ± 6.4	135
	100–150	10	104–149	134 ± 7.2	133.8
Slightly eroded, ploughland	0–50	10	108–147	130 ± 4.9	154
	50–100	10	69–150	100 ± 8.4	139.8
	100–150	10	64–130	99 ± 6.6	130.8
Strongly eroded, fallow	0–50	10	123–175	144 ± 5.1	145
	50–100	10	73–130	93 ± 6.7	119.2
	100–150	10	76–178	114 ± 6.8	92.6
<i>Priobye (1984–1987)</i>					
Non-eroded, virgin land	0–50	6	116–169	135 ± 6.8	149
	50–100	6	69–115	96 ± 8.5	120
	100–150	6	58–97	86 ± 6.5	109
Non-eroded, ploughland	0–50	6	116–164	136 ± 6.8	146
	50–100	6	78–127	90 ± 7.5	111
	100–150	6	73–132	86 ± 9.5	107
Slightly eroded, ploughland	0–50	14	91–149	123 ± 5.5	146
	50–100	14	60–126	87 ± 5.3	104
	100–150	14	53–137	85 ± 6.5	100
Slightly drifted ^b , ploughland	0–50	3	179–229	203 ± 1.5	160
	50–100	3	143–157	148 ± 4.5	115
	100–150	3	132–164	151 ± 9.7	125
<i>Kuznetsk Depression (1968–1979)</i>					
Non-eroded, virgin land	0–50	8	130–206	165 ± 9.3	206
	50–100	8	87–162	114 ± 9.4	155
	100–150	8	101–174	122 ± 8.7	143
Non-eroded, Ploughland	0–50	9	111–201	149 ± 9.2	192
	50–100	9	90–160	122 ± 6.3	154
	100–150	9	109–155	129 ± 5.5	143
Slightly eroded, ploughland	0–50	11	123–204	153 ± 8.0	151
	50–100	11	80–136	114 ± 5.0	143
	100–150	11	86–137	110 ± 4.2	138
Moderately eroded, ploughland	0–50	11	107–195	136 ± 9.2	156
	50–100	11	86–142	113 ± 4.8	141
	100–150	11	112–137	124 ± 3.1	135

(continued)

Table 7.3 (continued)

Degree of soil erosion, type of land use	Layer (cm)	Statistical parameters			FWC (mm)
		<i>n</i>	Range (mm)	$M \pm m$ (mm)	
Strongly eroded, ploughland	0–50	9	104–176	131 ± 7.7	153
	50–100	9	75–133	106 ± 5.7	151
	100–150	9	88–124	106 ± 3.6	135
Slightly drifted, ploughland	0–50	9	124–201	165 ± 12.4	177
	50–100	9	110–178	136 ± 7.7	143
	100–150	9	145–181	164 ± 4.5	146

^aMoisture reserves are the moisture content in the soil layers before and after snowmelt; ^bSlightly drifted soils are located in the lower position (bottom part) of catenae, where the solid component of the surface meltwater runoff accumulates; FWC is the field water capacity

Table 7.4 Spring moisture reserves^a in the profile of Chernozems before snowmelt (before the line) and their increase after snow melting (beyond the line) (mm)

Layer (cm)	Degree of soil erosion					
	Non-eroded		Slightly eroded	Moderately eroded	Strongly eroded	Slightly drifted
	Virgin land	Plough land				
<i>Predsalairye</i>						
	<i>n-10</i>	<i>n-4</i>	<i>n-11</i>	–	<i>n-11</i>	–
0–50	214/27	189/0	166/36	–	176/32	–
50–100	158/21	127/-2	102/2	–	144/51	–
100–150	146/12	126/0	116/17	–	141/27	–
0–150	518/60	442/-2	384/55	–	461/110	–
<i>Priobye</i>						
	<i>n-6</i>	<i>n-6</i>	<i>n-14</i>	<i>n-3</i>	–	<i>n-3</i>
0–50	148/13	165/29	165/40	142/46	–	189/-14
50–100	95/-1	100/10	98/11	78/25	–	133/-15
100–150	82/-4	91/-3	88/3	69/4	–	138/-13
0–150	325/8	356/36	349/54	289/75	–	460/42
<i>Kuznetsk depression</i>						
	<i>n-8</i>	<i>n-9</i>	<i>n-11</i>	<i>n-11</i>	<i>n-9</i>	<i>n-9</i>
0–50	188/23	197/48	153/0	151/15	137/6	182/17
50–100	109/-5	138/16	107/-7	102/-11	106/0	142/6
100–150	117/-5	127/-2	109/-1	108/-16	101/-5	148/-16
0–150	414/13	462/62	369/-8	361/-8	344/1	472/7

^aMoisture reserves are the moisture content in the soil layers before and after snowmelt

autumn and its rather deep and severe freezing during the cold period of the hydrological year have a negative impact on the addition of

moisture during the snow-melting period. In the southeast of Western Siberia, by the end of snowmelt, the soil thaws to a maximum of

40 cm. Deeper than this, regardless of the winter snowfall, soils are frozen and are not able to absorb more water. Therefore, the increase in the moisture reserve when the snow finishes melting depends on the initial soil moisture in the autumn and the depth of the thawed layer in the spring.

The eroded soils of Predsairye, Priobye and the Kuznetsk Depression have different water storage capacities in the pre-winter period. The highest values of this indicator (about 100 mm) were observed on the strongly eroded Chernozems in the Kuznetsk Depression, the smallest (51 mm) on the slightly eroded Chernozems of Predsairye. When the snow melts, the soil on the slopes thaws to a depth of no more than 40 cm: thus, even with a significant spring moisture deficit, they are not able to absorb the amount of meltwater contained in the snow. Consequently, by using data on the pre-winter moisture in the soil profile, it was possible to predict both the accumulation of meltwater in the soil profile and the volume of surface runoff of this water.

7.4.4 Depth of Soil Freezing

The term ‘permafrost regime’ includes the rate of freezing and thawing of the profile, the depth of penetration and the duration of subzero temperatures at different depths, the depth of the seasonal layer of permafrost, etc. All these parameters depend mainly on three factors mentioned above: the air temperature in the cold period, the amount of solid precipitation (height of snow cover) and the degree of soil moisture before freezing. In Western Siberia, snow cover is the most powerful factor. According to Rikhter (1945), the snow layer breaks the heat exchange in the soil-air system into two independent fragments: the stronger the snow cover, the more the air and soil begin to differ in temperature. Therefore, the effect of snow cover on the temperature regime of Chernozems can vary from year to year. In very low- and low-snow

hydrological years, relatively shallow snow cover is formed in November, mainly until the end of the month. As a result, subzero temperatures actively penetrate deep into the soil. During this period, the temperature of 0 °C penetrates into the soils of Predsairye and the Kuznetsk Depression to a depth of 30–40 cm. Significant precipitation in the form of snow in snowy and very snowy years in November (>40 mm), even against the background of significant subzero air temperatures, has a positive effect on soil freezing in the southeast of Western Siberia (freezing up to 20 cm). In both Predsairye and the Kuznetsk Depression, in December, there is slightly less precipitation than in November (Table 7.2) and the atmosphere during this period cools much more strongly, which stimulates the active penetration of subzero temperatures deep into the soil. As a result, at the end of December, at a depth of 80 cm, the temperature is fixed at –2 °C, and in the 0–10 cm layer it drops to –6 °C. In January and February, in very low-snow hydrological years in Predsairye and the Kuznetsk Depression, an extremely small amount of snow falls against the background of a very low air temperature. During these 2 months, 20–30 mm of precipitation falls, and the average air temperature is –18 °C. Such extreme climatic conditions can greatly affect the depth of soil freezing. At the end of February, the zero isotherms are established at a depth of 180 cm, and at a depth of 70 cm, the temperature can drop to –2 °C. The air temperature in March rises in comparison with February by about –10 °C, so soil freezing slows down significantly. The rate of soil freezing in very low-snow hydrological years in November reaches 2 cm per day, and in March it slows down to 0.6 cm per day. In April, the air temperature rises above zero (>1 °C) and the snowless soil begins to thaw quickly.

The soil temperature regime in Priobye differs from the soils of the forest steppe. In the Chernozems of the forest steppe, the freezing depth is closely related to the air temperature and the height of the snow cover, while in the

Chernozems of Priobye, this is not the case. This can probably be explained by the extreme dryness of the soil in the pre-winter period (Table 7.3). During dry periods of vegetation, amount of moisture in the 0–50 cm layer in the pre-winter period are even lower than the capillary rupture humidity, which indicates that the soil profile dries out to a very high degree. It is known that the drier the soil profile, the deeper the subzero temperatures penetrate. Consequently, the significant aridity of Priobye soil means that, regardless of the snowiness of the winter period, the soil freezes deeply and strongly under the influence of subzero air temperatures.

The soils of the Kuznetsk Depression are characterised by the maximum freezing depth among the Chernozems of the southeast of Western Siberia (Table 7.5). In very low-snow winters, these soils freeze to almost 210 cm. The

depth of their freezing in very snowy years is 2.7 times less than in very low-snow years. In very low-snow winters, the soils of Predsairye are characterised by a smaller depth of freezing (by about 1/3) compared to the Kuznetsk Depression. However, in very high-snow winters, the soils of Predsairye freeze deeper than the Kuznetsk Depression.

The soil temperature in the key plots during the cold period was studied based on the time/temperature charts that were built in the autumn. We obtained reliable temperature data from small (coin-sized) electronic temperature recorders (Thermochron iButton Device 2019). To study the temperature regime of the arable soil layer (soil surface and at depths of 5, 10, 15 and 20 cm) the temperature recorders were placed in individual protective pots and then placed in the soil. To check the temperature of the subsurface layers, several holes with a

Table 7.5 The depth of freezing of Chernozems in different hydrological years

Characteristics of hydrological years	n	Statistical parameters			
		Range (cm)	$M \pm m$ (cm)	δ (cm)	V (%)
<i>Predsairye (1950–2017)</i>					
Very low-snow	3	150–180	163 ± 8.8	15.3	9
Low-snow	13	110–171	145 ± 5.6	20.2	14
Normal-snow	25	15–181	129 ± 8.1	40.5	31
High-snow	12	70–160	120 ± 10.5	36.4	30
Very high-snow	12	25–175	112 ± 13.6	47.2	42
Extremely high-snow	2	20–30	25	–	–
<i>Priobye (1941–1990)</i>					
Very low-snow	8	73–240	134 ± 23.4	66.2	49
Low-snow	13	53–246	143 ± 19.6	70.5	49
Normal-snow	22	52–260	184 ± 12.2	56.9	31
High-snow	5	76–227	157 ± 28.9	64.7	41
Very high-snow	1	–	213	–	–
<i>Kuznetsk depression (1950–1987)</i>					
Very low-snow	7	139–286	207 ± 19.8	54.3	26
Low-snow	11	40–248	180 ± 17.0	56.4	31
Normal-snow	6	125–215	154 ± 14.1	36.4	22
High-snow	7	30–132	93 ± 12.7	33.4	36
Very high-snow	3	10–125	75 ± 26.6	46.0	61

diameter of 5 cm were drilled at a depth of 40, 60, 80, 100, 120, 140 and 160 cm. The temperature recorders were fixed on the lower end of plastic pipes with a metal cover and then placed on the bottom of the hole in close contact with the soil. A wooden bar was inserted into the plastic pipe, isolating the thermostat from water and heat.

The temperature recorders were initially programmed to measure the temperature at regular intervals. The rate of frost penetration into the soils of Western Siberia, depending on weather conditions, can range from 1 to 6 cm per day. It is very important to determine when frost starts to penetrate the soil. But it is more important to record the moment the soil begins to thaw in the spring and to monitor the daily dynamics of soil temperatures at different depths during snow melting. The daily range of air temperatures during snow melting in this region is large, from +16 °C during the day, to – 15 °C at night (Tanasienko 2003). Therefore, the recorders were programmed with a measurement interval of 3 h. Free gravitational moisture in the soil freezes at 0 °C. Therefore, we considered the depth of soil freezing equal to the depth of the soil profile with a temperature of 0 °C and below.

7.4.5 Weather Conditions During the Snowmelt Period

In the south of Western Siberia, during the snow melting, high air temperatures and direct solar radiation have the greatest impact. According to Rutkovskaya (1962), spring weather conditions in the south of Western Siberia are divided into three types: (1) radiative, (2) advective and (3) mixed radiative–advective and advective–radiative. The duration of snowmelt, the intensity of runoff and the removal of soil material on the slopes depends on the type of spring weather.

Radiative conditions are characterised by sunny, calm weather with a rapid increase in daytime air temperatures above 0 °C and the persistence of subzero temperatures at night, plus, as a rule, an increase (during snow melting) in the intra-soil ice layer. This type of spring

weather occurs in 30% of cases. Snowmelt of this kind is characterised by high moduli and the largest volumes of surface drainage and soil leaching over a fairly short period (4–7 days).

The advective type of spring weather is characterised by a gradual increase in daytime temperatures above 0 °C and the persistence of subzero night-time temperatures. The predominance of low and high cloudiness in the daytime determines the long period of snow melting (18–27 days). With such a prolonged melting of snow, the intra-soil ice barrier is destroyed, and by the end of the melting, the temperature of the soil profile has risen above 0 °C (Tanasienko and Chumbaev 2010). Surface runoff under such conditions is minor. This type of spring weather occurs in 15–20% of cases.

The radiative–advective and advective–radiative subtypes of spring weather are characterised by a gradual increase in daytime air temperatures above 0 °C and night temperatures well below zero. A return of subzero daytime air temperatures is also possible. This can last 1–5 days, and can be repeated several times during the same spring period. These subtypes of spring snowmelt are usually cloudy or windy with clear spells. The frequency of this weather is 30–35%. This subtype of snowmelt lasts 10–14 days with a high runoff coefficient (0.55–0.65).

7.5 Field Measurements of the Main Components of the Soil Erosion Process

The number of days with an active meltwater surface runoff, the volume of meltwater runoff and the daytime air temperature during the period of snow melting, as well as the amount of the solid phase of the soil washed away by meltwater, are the most important indicators when studying the process of soil erosion and are determined directly in the field.

Meltwater surface runoff was studied both at temporary runoff sites (average area of 500 m²) and at elementary catchments (depressions) in a small area (10–15 ha) (Fig. 7.5). In the study of the snowmelt runoff and the washout of the solid



Fig. 7.5 Surface runoff on short-term runoff section (a) and on elementary catchments (b)

phase of the soil, runoff sections of different lengths were used (from 20–30 m to 300–400 m and more). Studies by Orlov (1977) showed that in the forest steppe zone of Western Siberia on convex–concave slopes, the optimal length of runoff sections is 100–150 m. The advantage of the runoff section is that the amount of surface meltwater can be determined with great confidence, but quite often the loss of the solid component of the flow is distorted.

The runoff depth is the amount of water flowing from an elementary catchment for any period, expressed in a layer (mm) which is evenly distributed over the areas (Taratunin 2008). An area of 1 ha with a runoff depth of 1 mm holds 10,000 L of water. To determine the depth of snowmelt runoff, the runoff intensity was measured, i.e. the amount of water (mm) generated by melting snow per unit of time. The flow rate was expressed in litres per second per ha ($l\ s^{-1}\ ha^{-1}$).

With the help of a measuring cylinder (1 L) and a stopwatch, the depth of the snowmelt runoff was determined at intervals of one hour. First, the intensity of the runoff was determined by recording the time (in seconds) required to fill the measuring cylinder with meltwater. The runoff intensity ($l\ s^{-1}$) was multiplied by the corresponding coefficients to calculate the runoff depth per hectare (Chumbaev and Tanasienko 2016).

On an elementary watershed of 10–15 ha, the volume of snowmelt runoff was determined in

the lower part of the narrow drainage. Knowing the depth, width and speed of the water flow per time unit in a narrow open drainage and elementary catchment, the intensity of the snowmelt runoff was calculated. The depth and width of the water flow in a narrow open drain was measured using a ruler or a metal snow scale. Traditionally, water flow can be determined by placing light pieces of material (such as foam) in a water stream and determining the time it takes to travel a certain distance through the stream of water. All measurements were repeated 3–5 times.

The day with the first signs of meltwater surface runoff (a drop in the drainage level or a water layer on the soil surface in the lower part of the runoff trench) is marked as the first day of active surface runoff. The last day of the snowmelt water surface runoff is considered the day when the meltwater on the slope studied is not observed either in the areas of temporary runoff or in the runoff depression.

The climate in the southeast of Western Siberia determined the temperature regime of the Chernozem soils which were studied, and especially the freezing and thawing processes, which largely determine the course of elementary soil formation, the nature of the moisture behaviour in the soil profile, and other processes. In particular, as suggested by Khmelev and Tanasienko (1983), intense freezing and late thawing of Chernozems negatively affects their water regime. Since the moisture capacity of these soils is very low due to the presence of permafrost in

Table 7.6 Snow water equivalent and surface runoff in different hydrological years

Characteristics of hydrological years	n	Snow water equivalent		Surface runoff		Runoff coefficient
		Range (mm)	M (mm)	Range (mm)	M (mm)	
<i>Predsalairye</i>						
Very low-snow	1	–	56	–	45	0.80
Low-snow	5	1–67	39	1–74	41	0.51
Normal-snow	13	96–124	111	2–87	55	0.50
High-snow	13	135–155	146	19–111	68	0.46
Very high-snow	7	160–185	172	1–130	71	0.41
Extremely high-snow	2	187–210	198	89–142	115	0.58
<i>Priobye</i>						
Very low-snow	4	18–71	50	2–45	27	0.54
Low-snow	3	78–87	84	35–47	41	0.49
Normal-snow	4	92–99	96	48–70	60	0.62
High-snow	3	109–119	113	58–78	69	0.61
Very high-snow	3	129–149	138	60–91	73	0.53
<i>Kuznetsk depression</i>						
Very low-snow	1	–	58	–	0	0
Low-snow	2	–	60	–	33	0.55
Normal-snow	1	–	90	–	58	0.64
High-snow	1	–	115	–	75	0.65
Very high-snow	2	105–283	194	74–191	132	0.68

the soil profile during snow melting, large losses of meltwater with surface runoff occur on highly dissected terrain (Table 7.6).

The reason for the high meltwater surface runoff in very low-snow winters in Predsalairye, and especially in Priobye, is the low permeability of frozen soils due to their very deep and strong freezing (160–207 cm). Previous studies have shown that for frozen or unfrozen soils, permeability decreased with increasing soil moisture, but the effect of moisture on the infiltration rate decreased over time (Bodman and Colman 1944; Kuznik 1962). As the freezing depth decreases, the meltwater surface runoff gradually decreases.

Removal of the solid phase of the soil depends, firstly, on the permeability of the soil during snow melting, and secondly, on the intensity of snow melting. In the Chernozems of southeast Western Siberia, when the snow starts

to melt, large reserves of cold accumulate (Tanasienko 2003); therefore, these soils are practically waterproof. Moreover, during the melting of snow at night, frosts often reach -10°C , therefore the soil which has thawed during the day freezes again at night, making it impossible for the soil to fully absorb the meltwater.

The intensity of the erosion processes during snow melting characterises the amount of the soil solid phase in the runoff from the upper soil layer of the slopes. According to the studies by Maltsev and Yermolaev (2020), in the basins of small rivers in the European territory of Russia, erosion from melting snow is only $260 \text{ kg/ha}^{-1} \text{ year}^{-1}$. Our long-term studies have shown that from 0 to 150 kg/ha^{-1} of soil material is removed from non-eroded arable soils (Tanasienko 2003). This insignificant removal of the solid phase of

Chernozems with a significant snowmelt water surface runoff can be explained by the location of these soils in combined watersheds, where the surface runoff begins and therefore still has a low kinetic energy.

Slightly eroded soils are located below the watershed line, where the slope gradient is 3° . With an increase in the gradient in low-snow winters, the surface runoff on slightly eroded Chernozems does not exceed 40 mm (Table 7.6) with the washout of the soil material 2 t/ha^{-1} . In Priobye and the Kuznetsk Depression, in very low-snow and low-snow hydrological years, the removal of the soil solid phase on slightly eroded Chernozem ranges between 0 and 4 t/ha^{-1} , which is taken as the permissible value of erosion (Zaslavsky 1979). An increase in meltwater by 15–20 mm in normal-snow years significantly affects the amount of soil material removed by meltwater flows (Table 7.7). After high-snow winters, a large snowmelt water surface runoff was observed, which, leads to a significant removal of the solid component ($5\text{--}10 \text{ t/ha}^{-1}$). Its value for weakly eroded Chernozems in Predsairye, Priobye and the Kuznetsk Depression was diagnosed as average. In a very high-snow years, depending on the weather conditions during the snow melting, surface runoff can reach enormous values (up to 190 mm). Snow melting can occur in a relatively short time, which

predetermines a large daily flow rate, approaching a strong removal of the solid phase of weakly eroded Chernozems which are resistant to erosion (up to 20 t/ha^{-1}).

The question is: How can the annual loss of the soil solid phase be estimated depending on the snowiness of the hydrological year? If we take into account that, on average, 1 mm of the humus horizon of the soils studied has a mass of $9.0\text{--}10.4 \text{ t/ha}^{-1}$ (Khmelev and Tanasienko 2009), then soil loss in very low-snow, low-snow and normal-snow years does not reach even 0.5 mm. In high-snow and very high-snow years, the loss of the solid phase of the soil increases sharply and ranges from 4.5 to almost 20 t/ha^{-1} , i.e. fluctuates between 0.4 and 1.9 mm per year. Consequently, if in the very low-, low-, normal- and high-snow years (the share of which in the long-term climate cycle is more than 60%), the amount of washable arable soils is determined by their resistance to erosion, then in the very high-snow years the main factors of soil erosion are weather conditions during snowmelt and soil tillage. At this rate of solid-phase removal, in the absence of anti-erosion measures, these slopes can lose 20–22 cm in 100 years, which corresponds to half of the humus horizon.

The temperature of the meltwater directly affects the rate of soil thawing and thus these soils' ability to absorb meltwater. The higher the

Table 7.7 Loss of soil solid phase by meltwaters from slightly eroded Chernozems in different hydrological years

Characteristics of hydrological years	Predsairye			Priobye			Kuznetsk Depression		
	<i>n</i>	Range (t/ha)	<i>M</i> (t/ha)	<i>n</i>	Range (t/ha)	<i>M</i> (t/ha)	<i>n</i>	Range (t/ha)	<i>M</i> (t/ha)
Very low-snow	1	–	4.1	4	0–1.4	0.9	1	–	0
Low-snow	5	0.2–3.3	2.9	3	0.8–3.0	1.9	1	–	0.8
Normal-snow	13	1.0–28.3	7.6	4	1.4–6.1	3.9	1	–	1.4
High-snow	13	0.6–12.1	5.0	4	8.5–11.9	10.2	2	1.3–3.9	2.6
Very high-snow	7	0.3–28.8	8.6	4	1.4–14.2	9.9	2	9.0–9.7	9.3
Extremely high-snow	2	1.1–1.5	1.3	–	–	–	–	–	–

temperature of the meltwater, the greater the thickness of the thawed arable layer, which is in a so-called thixotropic (liquid) state due to oversaturation with meltwater. The temperature was recorded by an autonomous 'Termochron' thermometer while simultaneously measuring the hourly volume of meltwater runoff in the lower parts of the temporary runoff areas and runoff gutters.

7.6 Assessment of Soil Erosion Effects

Erosion is the process of destruction of rocks and soils by water flow, the movement of products of destruction and their re-deposition. To assess the consequences of soil erosion, it is necessary to determine the amount of soil material captured by meltwater from the fields. For this, the hourly turbidity of the water flow at the bottom of the drain gutter was determined, accompanied by checks at temporary runoff sites. The turbidity of the water flow characterises a decrease in the transparency of the water due to the presence of fine inorganic and organic suspensions. The turbidity of the meltwater was determined in the laboratory on an hourly basis using field samples taken from the experimental runoff plot using a 1 L measuring cylinder. The snowmelt water from a cylinder was transferred into a plastic jar and delivered to the laboratory. On the same day, the soil solid phase was separated from the liquid by filtration or centrifugation. For further analysis, the filtered water was placed in a refrigerator. The filtrate was weighed and the amount of solid phase in the soil was determined per litre of snowmelt water sample. The turbidity was evaluated in grammes per litre of water (g/l). Then, the amount of solid sediment washed off by meltwater per hour and per day was calculated. At the end of the snowmelt period, the data were summarised and the amount of solid phase displaced by meltwater over the entire period of snow melting was calculated.

In Western Siberia, erosion processes are intensifying due to the long-term use of

Chernozems on the slopes nowadays. The conditions of the erosion process revealed that neither snow water nor abundant precipitation indicates the severity of erosion. They only create the conditions for the formation of surface runoff of meltwater and rainwater. The intensity of the development of erosion processes can be judged by the amount of the soil solid phase in the snowmelt water runoff and the humus content in them. To assess the intensity of soil washout, Zaslavsky (1979) proposed the following scale, based on the annual removal of the solid phase of the soil: low washout—annual soil loss of 2–5 t/ha⁻¹; medium—5–10 t/ha⁻¹; strong—10–20 t/ha⁻¹ and very strong—average annual soil losses exceeding 20 t/ha⁻¹.

A common criterion for assessing the susceptibility of soil to erosion is the scale of the decrease in the thickness of the upper (humus) horizons of the Chernozem profile. In assessing the degree of soil erosion in Western Siberia, virgin soil was taken as the reference. In these soils, the upper, most humified part of the profile consists of a complete set of genetic horizons: Ah – AhB – B. If the thickness of Ahp + Ah horizons of arable soils on the slopes is less than the same reference horizon, then such soils are considered to be slightly eroded. When the AhB horizon partially enters the arable layer, these soils are moderately eroded. When the arable layer consists of a mixture of Ah, AhB and B horizons, this means that the initial horizon Ah has been eroded and such soils are strongly eroded.

Solid runoff products contain various amounts of the biogenic elements involved in the formation of soil structure and provide nutrients for cultivated crops, with organic carbon nitrogen, phosphorus, sodium, calcium and magnesium being the major nutrients (Coble et al. 2019; Hoffman et al. 2019; Wilson et al. 2019; Rasmussen et al. (2020). According to Nearing et al. (2017), soil erosion is the source of 80% of the total phosphorus and 73% of the total nitrogen in waterways in the United States. The same nutrients are present in the liquid phase of surface runoff. Their especially high content was

observed in the initial and final phases of runoff, when the time of contact of meltwater with the surface of slightly thawed soils is greatest.

In the light of climate warming, particularly in the cold regions, the export of dissolved organic carbon with the snowmelt has become a serious hazard (Meingast et al. 2020) to soil fertility. Almost every year, meltwater washes away a significant amount of water-soluble humus, mainly represented by mobile humic acids and calcium, which play an important role in soil formation, acting as organic and mineral adhesives (Kononova 1951). The amount lost increases many times with an increase in the equivalent of snow water and the amount of surface meltwater runoff. The washing out of soluble calcium and humus weakens the bonds between soil aggregates and creates the conditions for the subsequent separation of these aggregates by the surface waters when the kinetic energy of the flow is high.

The losses of water-soluble humus in various regions of Western Siberia in different hydrological years are given in Fig. 7.6. Naturally, the lowest losses of water-soluble carbon were observed in very low-snow winters. In the low-snow hydrological years with small losses of snowmelt waters, carbon removal from the soils of Priobye remains almost at the same level: twice the level as from the Predsairye soils. The latter have low losses of water-soluble carbon in the low-snow hydrological years, and a neutral reaction of the arable soil layer. In normal-snow hydrological years, carbon removal by snowmelt

waters in all regions of Western Siberia was almost the same.

In high-snow hydrological years, the removal of carbon by meltwater was almost the same in all regions of Western Siberia, while in very high-snow years, the difference increases significantly. Thus, for example, the removal of water-soluble carbon from the Chernozems of Predsairye and Priobye in a very high-snow year was 7 and 4 times greater, respectively, than in a low-snow year, and almost twice as high as in a high-snow year. Losses of organic carbon with meltwater from the Chernozems of the Kuznetsk Depression in very high-snow years were almost 1.5 times greater than in similar years in Priobye and Predsairye. This large removal of carbon can only be explained by the fact that the Chernozems of the Kuznetsk Depression contain a very large amount of carbon in the arable layer (5.5%), while in similar soils of Predsairye and Priobye, the content of this element is much lower (4.5 and 3.5%, respectively). In addition to the losses of water-soluble carbon, mineral nitrogen and phosphorus are lost concurrently. The nitrogen content in the meltwater was insignificant, however, given that, on average, 850 m³ of meltwater flows from the soil surface of Predsairye and the Kuznetsk Depression annually: the loss of this element reaches almost 2 kg/ha⁻¹ (Fig. 7.7).

The nitrogen content in the solid component of surface runoff was from 0.4 to 0.7% on average. A significant removal of nitrogen from the solid phase of the soil led to a sharp decrease

Fig. 7.6 Carbon loss from slightly eroded Chernozems with liquid and solid component of meltwater surface runoff in Predsairye (1), Priobye (2) and the Kuznetsk Depression (3) sites

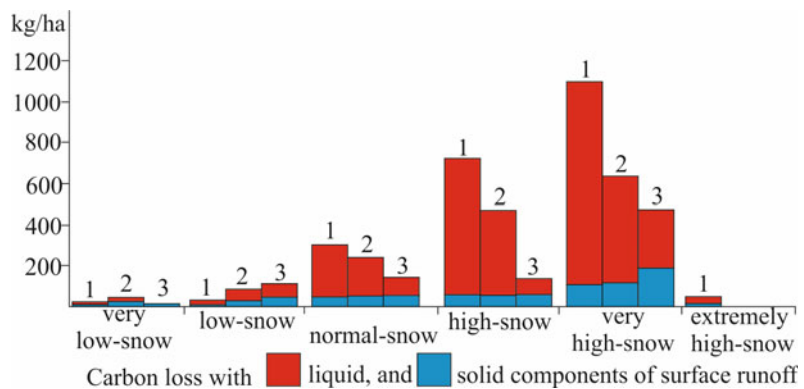
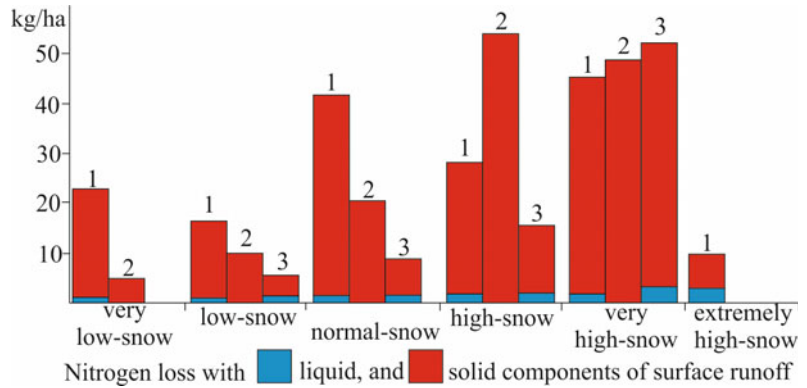


Fig. 7.7 Nitrogen loss from slightly eroded Chernozems with liquid and solid components of meltwater surface runoff in Predsairye (1), Priobye (2), Kuznetsk Depression (3)



in the amount in the profile of eroded soils. The nitrogen concentration in solid runoff was approximately 0.1–0.2% higher than in the arable layer of slightly eroded Chernozems (Tanasienko et al. 2016). Calculations show that in these soils the annual removal of nitrogen from solid runoff, depending on the snowiness of the hydrological year, was between 15–50 kg/ha⁻¹. Absolute nitrogen losses due to erosion during the period of intense anthropogenic impact on the soil of slopes reached 6.6 t/ha⁻¹ for slightly eroded Chernozems, 12.2 t/ha⁻¹ for moderately eroded Chernozems and 15.5 t/ha⁻¹ for strongly eroded Chernozems.

A decrease in the depth of the humus horizon under the influence of erosion processes led to a regular decrease in the content of total phosphorus. Since the maximum amount of biogenic elements is concentrated in the upper part of the humus layer, during the meltwater surface runoff, their losses occur together with the solid phase of the soil. As a rule, in very low-snow and low-snow hydrological years, the removal of gross phosphorus is low (no more than 5 kg/ha⁻¹ annually). Losses of phosphorus with a solid runoff of meltwater in normal-snow winters in the Kuznetsk site were almost three times greater than in low-snow winters. The annual loss of phosphorus in very high-snow years exceeded 30 kg/ha⁻¹, twice as much as after high-snow winters.

Our results showed that in the Chernozems at higher elevations, a general increase in moisture content can provide a periodic type of leaching

regime with more frequent soil moistening in early spring. Due to the significant surface runoff of snowmelt water, the accumulation of moisture in the slope is sharply reduced, which affects their water regime. In weakly eroded soils, the water regime formed is the non-leaching type, and in highly eroded soils, the so-called ‘drainage’ type. Due to the influx of moisture with surface runoff to the reclaimed soils, they periodically form a leaching water regime. The soils of the sloping territories of Predsairye, Priobye and the Kuznetsk Depression, regardless of the snowiness of a hydrological year, freeze for more than five months, which significantly limits the absorption of snowmelt water and stimulates significant surface runoff.

Almost all Chernozem soils of Western Siberia have been cultivated for a long time; therefore, they should be considered anthropogenically transformed soils. From 30 to 130 mm of meltwater, large areas of slopes are lost annually as a result of surface runoff, causing an intensive development of erosion processes, accompanied by the loss of 1–5 tonnes per ha of soil material, and lower amounts of humus, nitrogen and phosphorus.

7.7 Conclusions

- Due to the strong impact of erosion on the slope soils, the initially highly humified full-profile Chernozems were transformed into

moderately and slightly humified soils in the southeast of Western Siberia. On the slopes of any exposure, a decrease in the thickness of the humus horizon, which exceeds the level of natural variability, indicated the formation of eroded soils.

- Ploughing of virgin lands, unregulated surface runoff, flushing, transport and re-deposition of the soil solid phase by snowmelt flows (mostly) resulted in the formation of elementary soil areas that differ from those on the elevated lands. An increase in thickness in the lower positions of the landscape (slope paths) indicates the appearance of eroded (stratified) soils.
- The concentration of water-soluble humus and nutrients in meltwater depended both on the humus content in the depth of arable layer and on the amount of pre-winter moisture in the soil profile.
- The formation of a permanent snow cover late in the year promoted the free penetration of sub-zero temperatures into the soil profile, affecting the soil's water and temperature regime.
- Analysis of long-term data on the amount of solid precipitation falling in different hydrological years in Predsalaurye, Priobye and the Kuznetsk Depression showed that the significant amount of snow accumulating annually is sufficient to form surface runoff and erode the solid phase of arable slopes.
- The methods proposed for studying soil erosion during snowmelt are simple to use and do not require the use of expensive equipment in the field.

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