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GEOGRAPHICAL STUDIES No. 136

VERTICAL ZONALITY IN THE SOUTHERN KHANGAI MOUNTAINS (MONGOLIA)

WROCŁAW · WARSZAWA · KRAKÓW · GDAŃSK
ZAKŁAD NARODOWY IMIENIA OSSOLIŃSKICH
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ГЕОГРАФИЧЕСКИЕ ТРУДЫ

№ 136

ЗОНАЛЬНОСТЬ ЮЖНОГО СКЛОНА ХАНГАЯ (МОНГОЛИЯ)

РЕЗУЛЬТАТЫ ИССЛЕДОВАНИЙ МОНГОЛО-ПОЛЬСКОЙ ФИЗИКО-ГЕОГРАФИЧЕСКОЙ ЭКСПЕДИЦИИ

Т. I

ПОД РЕДАКЦИЕЙ

К. КЛИМЕКА И Л. СТАРКЕЛЯ

POLISH ACADEMY OF SCIENCES
INSTITUTE OF GEOGRAPHY AND SPATIAL ORGANIZATION

GEOGRAPHICAL STUDIES No. 136

VERTICAL ZONALITY IN THE SOUTHERN KHANGAI MOUNTAINS (MONGOLIA)

Result of the Polish-Mongolian Physico-Geographical Expedition

VOL. I

ELIGIUSZ BRZEŹNIAK, KAZIMIERZ KLIMEK, ALOJZY KOWALKOWSKI, RADNARIN LOMBORINCHEN, TADEUSZ NIEDŹWIEDŹ, ANNA PACYNA, KAZIMIERZ PĘKALA, LESZEK STARKEL, TADEUSZ ZIĘTARA

Edited by

KAZIMIERZ KLIMEK AND LESZEK STARKEL

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PREFACE

The scientific co-operation of the Institute of Geography and Spatial Organization. Polish Academy of Sciences, with the Institute of Geography and Geocryology, Mongolian Academy of Sciences, resulted in the Polish-Mongolian Physico-Geographical Expedition to the Khangai Mts (Mongolia) in the summer seasons of 1974 and 1975. The expedition was led by Kazimierz Klimek.

Important contributions were made by the expedition members to the understanding of natural environmental conditions prevailing in the southern part of the Khangai Mts, where few studies of this kind have as yet been made. Some of the results of geological, geomorphological, climatological, hydrological and pedological research have already been discussed. Most of the articles were published in the *Bulletin of the Polish Academy of Sciences,* 1975 and 1977. The present collection is the first out of a set of volumes of *Geographical Studies* devoted to selected physiographic problems in Mongolia.

This book to which numerous authors have contributed was elaborated editorially in the Department of Geomorphology and Hydrology of the Mountains and Uplands, Polish Academy of Sciences in Cracow. We would like to express our thanks to Dr. Sylwia Gilewska for the translation of the book and critical comments. Dr. Grace Claire-Dąbrowska and Prof. Dr. Tomasz Komornicki helped us in translation of two chapters (on vegetation and soils). We are indebted to Mrs. Maria Klimkowa for the drawing of figures and editorial assistance.

Kazimierz Klimek
Leszek Starkel

ПРЕДИСЛОВИЕ

В рамках научного сотрудничества между Институтом географии территориальной организации Польской академии наук и Институтом географии и геокриологии (мерзлотоведения) Академии наук Монгольской Народной Республики, в летних сезонах 1974 и 1975 годов в горах Хангая (Монголия) работала Монголо-польская физико-географическая экспедиция, руководителем которой был Казимеж Климек.

Участники экспедиции внесли значительный вклад в изучение природной среды слабо ещё исследованного южного склона Хангая. Некоторые результаты исследований в области геологии, геоморфологии, климатологии, гидрологии и почвоведения были опубликованы в нескольких статьях, главным образом в Бюллетене Польской академии наук за 1975 и 1977 гг. Настоящий том Географических трудов открывает серию публикаций, посвященных избранным проблемам физической географии Монголии.

Книга, в которой принимало участие много авторов, была подготовлена к печати в Отделе геоморфологии и гидрологии гор и возвышенностей Института географии и территориальной организации Польской академии наук в Кракове.

Мы выражаем нашу благодарность д-р Сильвии Гилевской за перевод книги и критические замечания, а также д-р Грейс Клер-Домбровской и проф. Томашу Коморницкому, которые перевели две главы — о растительности и почвах. Мы также очень обязаны магистру Марии Климковой за выполнение иллюстраций и помощь в подготовке книги к печати.

Казимеж Климек
Лешек Старкель

KAZIMIERZ KLIMEK

I. MAJOR PHYSICO-GEOGRAPHICAL FEATURES OF THE SOUTHERN SLOPE OF THE KHANGAI MOUNTAINS

The Khangai Mountains are lying at the boundaries between two large structural-orographical units; the mountains of East and South Siberia and the plains of Central Asia (Aleksandrovskaya *et al.* 1964).

The main Khangai ridge extends about 700 km southeastward. It reaches on average 3000—3500 m asl., the elevation of Otgontengri, the highest peak, is 4031 m asl. (fig. 1). The undulating South-Khangai Upland borders the Khangai Mountains from the south. It is characte-

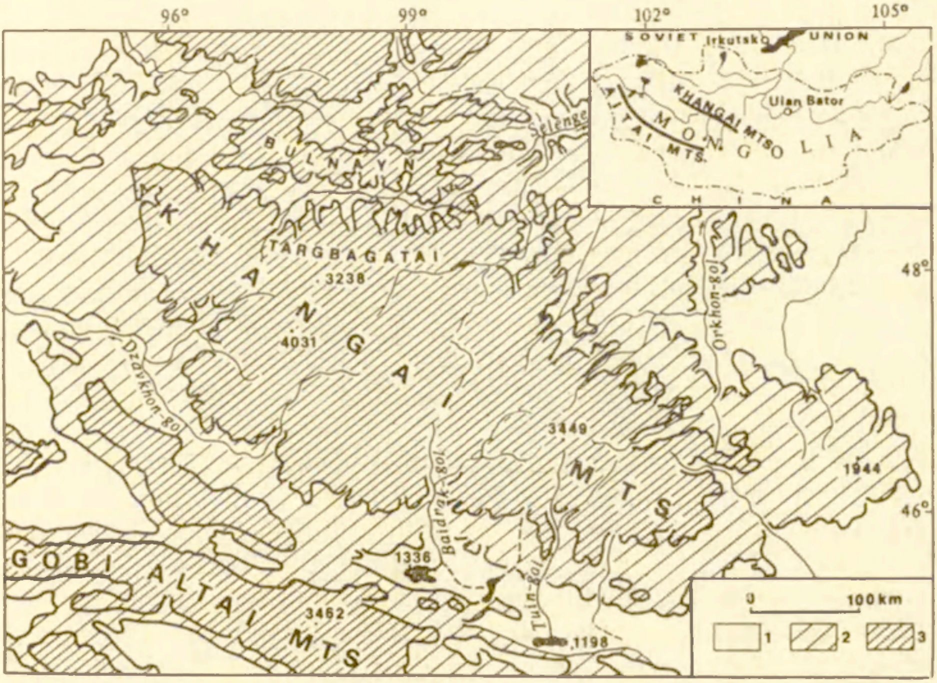

Fig. 1. Position of the Khangai Mts

1 — plains-plateaus and rift valley bottoms situated below 1000 m asl.; *2* — undulating plateaus and piedmonts extending between 1000 and 1500 m asl.; *3* — mountains and uplands rising above 2000 m asl.

rized by hills rising from 1500 m asl. at the margin of the Gobi to 2700 m asl. at the foot of the Khangai (fig. 2). Mountains of intermediate height form a wide belt on the northern side of the main Khangai ridge. Only the Tarbagatai range (3238 m asl.) and the Bulnayn range (2600 m asl.) attain higher elevations there.

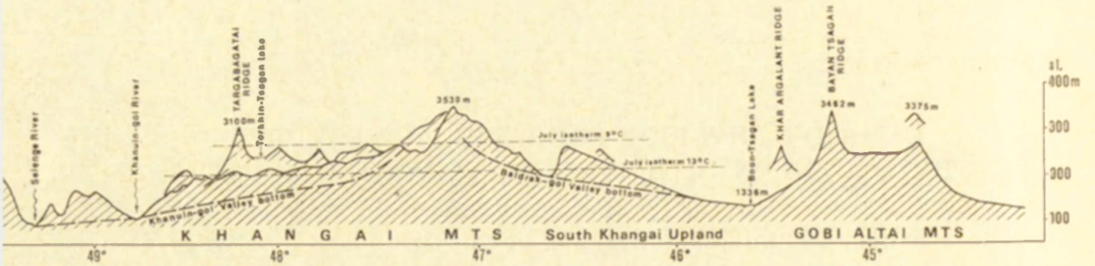

Fig. 2. Meridional profile of the Khangai Mts

The high relief of the southern Khangai reaching 1500 m over a distance of 40—60 km accounts for this clear orographic line in central Mongolia.

The Khangai Mountains, together with the northern piedmont comprise an elliptical tectonic block which is the western continuation of a larger unit referred to as the Khangai—Khentei (or Mongolian—Transbaikalian) fold system (fig. 3; Sonenschein 1973; Fillipova *et al.* 1973). This is a synclinorium which contains thick series of Devonian and Carboniferous sediments interbedded with continental volcanic and clastic rocks of Permian age. Numerous post-orogenic granite—diorite intrusions of either Lower or Upper Paleozoic age are associated with the much deformed sediments. The western part of the Khangai Block is also bordered for considerable distances by granitic rocks. The younger strata consist of basalt caps and flows. These are forming the summit areas and are also exposed in the valleys. The oldest basalt is dated at 7—12 million years. The younger basalt is Holocene (Kozhevnikov *et al.* 1970; Korina *et al.* 1974).

In the Khangai Block and in the southerly Baydrak Block numerous tectonic disturbances caused their subdivision into minor units. These dislocations were initiated at the turn of the Jurassic and the Cretaceous (Khasin 1973). The present tectonic relief of the area discussed dates from the Pliocene and the Quaternary (Khasin and Selivanov 1973). At that time the southern part of the planated Khangai Block was faulted and divided into smaller blocks, some of which rise to considerable heights. At present remnants of this undulating Cretaceous—Lower Tertiary planation surface are well preserved on the major watershed of the Khangai and in the South-Khangai Upland as well.

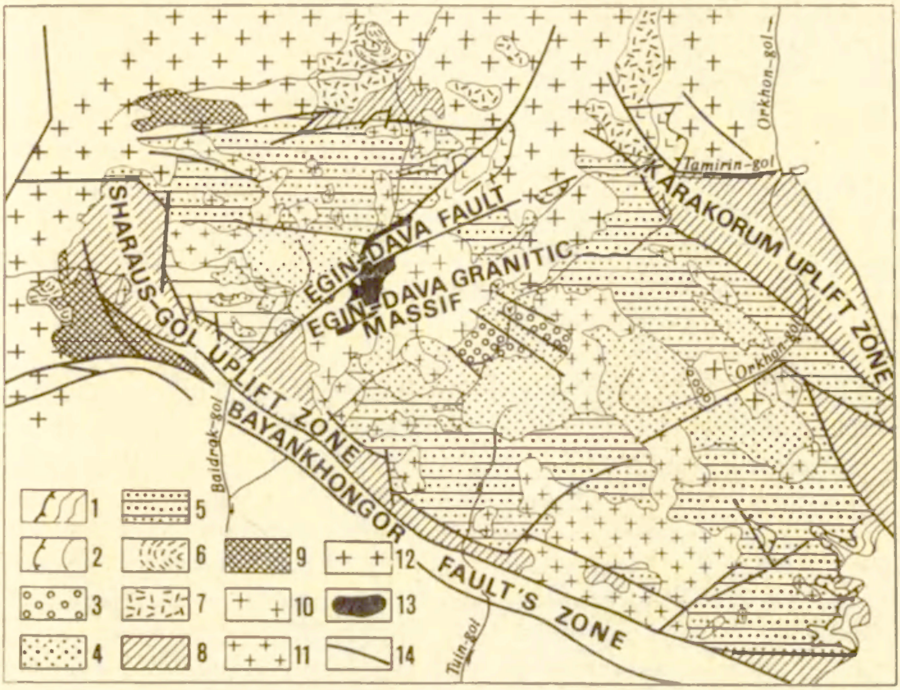

Fig. 3. Sketch showing the geological structure of the Khangai Block (according to L. P. Sonenschein 1973)

1 — Jurassic sedimentary series; *2* — Jurassic volcanic series; *3* — Permian deposits; *4* — upper part of Khangai series; *5* — lower part of Khangai series; *6* — Lower Carboniferous deposits; *7* — Paleozoic sedimentary-volcanic deposits; *8* — Lower Paleozoic sandy-shale deposits; *9* — Lower Cambrian deposits; *10* — Younger Mesozoic granitoides; *11* — Upper Paleozoic Khangai granodiorite; *12* — Upper Paleozoic granitic rocks; *13* — basalt cap; *14* — fault lines

The southern slope of the Khangai is a very clear structural-paleogeographical and geomorphological boundary. This is the expression of a major fracture of the earth's crust. At the foot of the Khangai this fracture separates two tectonic-structural units of varying geological history and of varying structure. The tectonic zone was still active during the Pleistocene, and evidence suggests that tectonic movements are taking place at the present day (Klimek and Rotnicki 1977).

During the Pliocene and the Quaternary the southern border of the Khangai Block was affected by intense tectonic movements which produced a high local relief attaining 1500 m. This is typical of the mountainous border of central Asia.

Many of the height differences are locally reduced by the South-Khangai Upland, and for this reason the southern slope of the Khangai is similar orographically to the borders of the Khentei Mts and the Khubsugul Mts. The high relief of the southern Khangai is responsible for the genetic varieties of the second-order relief features. The summit areas were fashioned by glacial and cryogenic processes. In the intermediate part fluvial processes were dominant, whereas fluvial and eolian sedimentation took place at the foot of the mountains.

Because of the "massive" appearance and high relief the southern slope of the Khangai reflects closely the boundary between South Siberia and Central Asia. This boundary is expressed in changes of the climatical, hydrological, soil and geobotanical conditions. Within this boundary zone the mountainous landscapes of South Siberia merge with the plains—plateau landscapes of the Gobi.

The Khangai lies within the temperate latitude climatic zone. The southern slope of the Khangai is the boundary between the cool continental mountain climates (with the characteristic belt-like arrangement of thermal and moisture conditions) and the extremely continental temperate warm climates of the Gobi (with zonal variations of thermal and moisture conditions). At the foot of the mountains the continentality of climate is reflected chiefly in the large diurnal and annual temperature ranges, in the low air humidity and the low amounts of precipitation (240—280 mm) concentrated in the short summer (comp. chapter IV). The high and contrasting relief of the Khangai favours the occurrence of pronounced temperature inversions, so that severe climatic conditions are found on the southern mountain slope. In the mountain foreland high mean monthly temperatures are recorded (Galuut experiences temperatures above 19°C), but the mean annual air temperature here is −5.3°C. The cause are the low winter temperatures. In January temperatures fall to −32°C, the absolute minimum being −48°C.

The almost complete lack of snow and cloud cover during the long autumn—winter—spring season favours both radiation and cooling of the ground. The heat loss cannot be balanced by summer warmth, so that discontinuous permafrost occurs in this area (Gravis 1974). Permafrost here reaches its southernmost limit (46°30′ φ) in the northern hemisphere. Patches of permafrost are found on north-facing slopes, and the cover of permafrost becomes more complete at higher altitudes. Permafrost also underlies valley floors and tectonic basins where the presence of frozen ground is controlled by the presence of ground water in the loose lacustrine and alluvial fills.

The severe climatic conditions are the major factors controlling hydrological conditions, soils, vegetation and morphogenetic processes occurring in the southern Khangai.

Along the central Khangai ridge runs the Continental Divide between the drainage basins of the Arctic Ocean and the endoreic areas of Central Asia. On the southern side of the central Khangai ridge the degree of continentality increases with increasing altitude. Precipitation amounts decline, while mean air temperatures and evaporation values increase (comp. chapter IV). In the summit area precipitation exceeds evaporation, so that a stream network can develop and surplus water is supplied to lower areas (Dauksza and Soja 1977). In the lower part of the southern Khangai, evaporation exceeds precipitation and there are surface water limitations. This area is drained by transitional streams which carry water gained on the ridges and feed it into the mountain foreland. In summer, thawing permafrost and rainwater floods cause locally greater ground moisture.

In the southern Khangai, soil formation was influenced by changes of both climate and plant communities with time. These factors were complicated by altitude, slope aspect and type of bedrock. Consequently there occurs a great variety of soil types. These include primitive brown soils (structure ground), mountain brown soils and mountain chernozem affected by congelifluction as well as dark chestnut soils. Light chestnut soils are found at the foot of the mountains (comp. chapter VI).

The southern slope of the Khangai is the boundary zone between two geobotanical provinces: the Euro-Asiatic province and the Central Asiatic province (Grubov 1955) referred to as the mountainous forest-steppe "Khangai" province and the desert-steppe "Gobi" province (Yunatov 1974). Patches of a boreal vegetation with *Larix sibirica* and *Pinus sibirica* which occur commonly in the northern Khangai here reach their southernmost limits. In the southern foreland of the Khangai steppes are widespread. These belong to the Mongolian—Gobi formation with *Artemisia* and *Stipa* (Likicheva 1964).

Because of the marked continentality of climate and high local relief (to 1500 m) a continental variety of the altitudinal zones of vegetation here is found (comp. chapter VIII). A characteristic feature of the southern slopes of both the Khangai Mts and the Khentei Mts is the occurrence of a forest—steppe belt. At the same altitude there occur either patches of larch forests or dry mountain steppes dependent on the orientation of slopes (either facing north or south).

In the Khangai the varying tectonic and paleoclimatic conditions are responsible for the production of various assemblages of landforms, sediments and weathering products being different from the present ones. At the present time there is a tendency for the inherited landscapes to become adjusted by exogenic processes to the actual conditions.

Because of marked height differences such elements of the geographical environment as climate, water and plant cover tend to change rapidly with increasing elevations. Typical high-mountainous tundra landscapes with perpetual snow patches thus may occur within sight with semi-desert landscapes.

KAZIMIERZ KLIMEK, RADNARIN LOMBORINCHEN, LESZEK STARKEL

II. HISTORICAL REVIEW OF PHYSICO-GEOGRAPHICAL INVESTIGATIONS IN THE KHANGAI MOUNTAINS

The wide mountain valleys with good pastures easy of access may explain why the nomads have lived in the Khangai from early times, although climatic conditions have been severe. The Khangai has been open to invasions of the Huns, the Turks and the Mongols (Gumilov 1960, 1967). To the east of the Khangai Karakorum the capital of the ancient Mongolian imperium was lying. The early nomadic tribes knew very well the physique of this mountain massif. At present herdsmen also know where to build camps within reach of the water and pastures having optimum climatic conditions.

The first information about the physical features of this area was provided, during the thirteenth century, by Marco Polo, Piano Carpini and other explorers, but the systematic geographical investigations of Mongolia (and of the Khangai Mts) began no earlier than in the nineteenth century.

PREVIOUS INVESTIGATIONS OF THE NATURAL ENVIRONMENT

In the second half of the nineteenth century the Russian Geographical Society began a systematic survey of the natural environment of Central Asia. Research was also extended to Mongolia. The first explorers to undertake physico-geopraphical problems of the Khangai were M. V. Pevtsov, N. M. Przewalski, G. N. Potanin, J. G. Gronno and P. K. Kozlov. These drew attention to the present belt-like distribution of environmental conditions and to the occurrence of former valley glaciations and lake level oscillations. The history of vegetation in this area was also studied.

Intense physico-geographical research began in 1921 with the creation of both the Mongolian People's Republic and the Mongolian Committee of Science, to be followed by the foundation of the first University in this country (1942). During this time-span many Soviet and joint Soviet—Mongolian expeditions were working in different parts of Mongolia and made notable geological and biological contributions. It is not possible to mention here all names of those who completed maps, monographs, reports and other contributions on the Khangai. Results of work were synthesized by Murzayev on physical geography (1952), by Kuznetsov on hydrology (1968), by Marinov and collaborators on geology (1973), by Selivanov on geomorphology (1972), by Gravis and collaborators on geocryology (1974), by Badarch on climatology (1971) and by Yunatov on the vegetation cover (1950). These syntheses provide general information about the evolution of the Khangai Mts and their geographical environment as well.

In the present study, of particular significance were the outstanding investigations of the vertical distribution patterns of natural phenomena, especially the works by Gravis on permafrost (1974) and by Karamysheva and Banzragch on vegetation (1977).

By about 1950 the need for more detailed study became apparent, and research of small areas was initiated. The development of physico-geographical investigations was connected with the foundation of the Mongolian Academy of Sciences. At the same time the demand arose for both recognition and evaluation of the natural resources for economic purposes. Foreign scientists are being invited to act in co-operation with the young Institute of Geography, Mongolian Academy of Sciences.

In the past two decades two international expeditions were undertaken to carry out complex detailed studies of small areas situated in the southern Khangai. In 1960 the German geographers sent their expedition to the Tsagan-Turutuin-gol drainage basin. Haase, Richter and Barthel from Leipzig have examined in detail the relief, soil, frost action processes and plant cover in a small valley adjacent to the Sant valley in which the Polish-Mongolian expedition was working in 1974—1975. The above mentioned authors dealt with the physico-geographical zoning of various phenomena on the northern and southern slopes of the Khangai. The belts recognized are as follows: the scree belt (Frostschuttzone), the alpine and subalpine belt, the mountain forest-steppe belt and the steppe belt (Haase *et al.* 1964). Various ecotopes were delimited within the slope catenas. Attention was drawn to the dependence of habitats upon permafrost occurrence which in turn appears to be controlled by slope aspect. A characteristic feature of mountains having a continental climate is slope asymmetry; the north-facing slope is smoother than the opposite one (Richter *et al.* 1963).

THE JOINT POLISH—MONGOLIAN EXPEDITION

The Polish-Mongolian physico-geographical expedition to the Khangai Mts was organized by the Department of Physical Geography, Institute of Geography and Spatial Organization, Polish Academy of Sciences in

Krakow acting in co-operation with the Institute of Geography and Geocryology, Mongolian Academy of Sciences in Ulan Bator. The expedition was led by Kazimierz Klimek. Principal work was made in the southern Khangai during two summer seasons (June—August), in 1974—1975. Studies were completed in April, 1976. A detailed report on the first Polish-Mongolian expedition to the Khangai was published in 1976 (Klimek *et al.* 1976).

This expedition included twenty Poles (17 scientific workers and 3 technicians) and three Mongols.

It was the aim of research (1) to gain a complex knowledge of the physico-geographical environment in the Uldzeitu-gol catchment, and (2) to outline the value of environmental resources for the economic development of this region.

The need for accurate data has resulted in the simultaneous studies of all components of the natural environment. Thus information on the interrelationships between these components was obtained in a short time. Various aspects of Quaternary geomorphology and paleography and of climatology, hydrology, pedology and botany have also been studied.

The study area extends from the central Khangai ridge down to the sub-Khangai tectonic basins and exceeds 2100 km^2. Since detailed topographical maps were lacking, in 1974 field work was carried out primarily along transects from the summit areas to the valley floors and along the major valley. Furthermore, climatic and hydrological phenomena and morphogenetic processes were currently measured in a few sites. The methods used in Mongolia have been worked out in the Polish Carpathians. In 1975 studies were concentrated mostly on the altitudinal zoning of various natural environmental factors.

In 1974—1975 four teams were working in different parts of the study area. These were as follows:

1. The main camp, together with a meteorological station and a stream gauging point was established at Mandal (at 2055 m asl.). This is within the steppe belt, where the Tsagan-Turutuin-gol river emerges from the Khangai Mts. Geomorphological, hydrological, pedological and botanical investigations were made in the camp's surroundings, in the Tsagan-Turutuin-gol valley and in the nearby Sant valley.

2. The Sant valley situated at heights of about 2100—2700 m asl. is 3 km^2 in area. It is carved into the mountain margin within the forest-steppe belt. Detailed investigations here were carried out under the direction of L. Starkel (in 1974) and of A. Kowalkowski (in 1975). An accurate topographical map produced during this time as well as observations completed in spring made it possible to solve the problem of both slope catenas and north-south asymmetry. Result of research will be published under separate cover in *Geographical Studies*, No. 137.

3. Within the belt of high-mountain meadows a meteorological station

(at 2650 m asl.) and a stream gauging point was established in the Olon-nuur valley in 1975. Geomorphological, hydrological, geobotanical and soil surveys were undertaken at higher levels in the Khangai and in the glacially modified valleys.

4. Within the dry steppe belt another meteorological station (at 1950 m asl.) and gauging point was established in the Bayan-Nuurin-khotgor Basin in 1974—1975. At the "Basin" station studies were concerned essentially with geomorphological problems of both the tectonic sub-Khangai basins and the South-Khangai Upland scarp.

The preliminary results of research are given in two manuscript volumes entitled "Report on research" (1975, 1976), and about 30 original contributions to the physico-geographical problems of the study areas were published in *Bulletin of the Polish Academy of Sciences*, 1975, 1977.

Within the field of geomorphology and paleogeography, Polish investigations support the views of other workers (Kozhevnikov *et al.* 1970) that extensive planation surfaces are associated with the summit areas of the Khangai (Starkel *et al.* 1975). The nature of the last glaciation and its extent in the Khangai was recognized by Klimek and Sugar (1975) and by Klimek (1977). In the mountain foreland large lakes are regarded as having been formed contemporaneously with the last Pleistocene valley glaciers (Klimek and Sugar 1975; Klimek and Rotnicki 1977). It was found that the mid-Holocene warming of climate was followed by a cooler and drier phase. This began about 2000 yrs B. P. (Kowalkowski *et al.* 1977). At higher levels extending above 2700 m asl. very active cryogenic processes produce cryoplanation terraces on the slopes (Starkel *et al.* 1975). The rate of present-day slope processes was determined in the marginal part of the mountains (Kowalkowski 1977; Pękala 1975; Pękala and Ziętara 1977) and during the interesting thaw season (Froehlich and Słupik 1977).

Within the field of climatology, summer air temperatures were correlated along a vertical profile in the mountains (Brzeźniak 1977), and diurnal variations of various climatic elements in summer were recognized at the foot of the mountains (Avirmid and Niedźwiedź 1975).

Research revealed that hydrological phenomena are distributed zonally and belt-like on the southern slope of the Khangai. It was found that there is a certain sequence of areas which differ in precipitation—evaporation ratios, i.e. from $P > E$ to $P < E$ (Dauksza and Soja 1977). The way in which water is received by the Tsagan-Turutuin-gol river has been determined and the amount of suspended load in summer measured (Froehlich and Sugar 1975; Froehlich *et al.* 1975). Both occurrence and disappearance of widespread icings was observed in spring (Froehlich and Słupik 1977).

Pedological studies revealed the presence of various soil types. Variations in both physical and chemical properties of the soils allowed different catenas to be delimited. These depend on slope aspect and altitude (Kowalkowski 1975, 1977; Kowalkowski and Lomborinchen 1975). Numerous catenas show relict features. These are the effect of past contrasting climates controlling the processes of denudation.

Geobotanical studies lead to the recognition of several hundred species of plants. Furthermore, different types of habitats were distinguished and related to other environmental factors in the marginal part of the mountains (Kowalkowski and Pacyna 1977).

In the published contributions various physico-geographical problems of the study areas are discussed. In the present volume a broader characteristics of the natural environment of the southern Khangai is attempted by taking height-dependent changes as well as continentality of climate fully into account.

KAZIMIERZ KLIMEK

III. RELIEF AND PALEOGEOGRAPHY OF THE SOUTHERN KHANGAI MOUNTAINS

Relief is important in controlling local variations of other environmental components within a mountain region because of high local relief and varying slope aspect.

RELIEF AND PALEOGEOGRAPHY OF THE SOUTHERN KHANGAI

Within the Uldzeitu-gol drainage basin the southern slope of the Khangai occupies a belt, 40—60 km wide. This area may be subdivided into two minor zones which owe their present nature to contrasting conditions of the past.

The internal zone, 10—20 km wide, comprises extensive flat surfaces which have been dissected by deep glacial cirques and U-shaped glacial troughs (fig. 4). These surfaces occur at three levels being constant in height (fig. 5). The highest surface forms a plateau on the major watershed, where it rises to about 3400 m asl. This plateau includes both flattenings, up to 5 km long and 1—1.5 km across, and conical residuals, up to 100 m high. In places the watershed plateau grades slightly into the lower, subwatershed plateau which reaches 3100—3300 m asl. This flat surface attains widths of 5—10 km. Deep valleys and glacier cirques have divided it into smaller isolated patches which may extend over an area 3 km by 5 km. At their outer margin, these flattenings give way to broad mountain ridges (fig. 4). In general, the sub-watershed surface has a relative relief of the order of 100—150 m.

The high valley-side flattenings (2650—2850 m asl.) are clearly developed along the Tsagan-Turutuin-gol valley (fig. 4). They are also found at the margin of the tectonic Bayan-Nuurin-khotgor Basin. These flattenings are most extensive in the interior of the mountains. Where the Barun Anag, Dund Anag and Dzun Anag valleys join, individual patches of the high valley-side flattenings may extend over an area 4 km by 8 km. This surface shows a clearly transverse slope towards the incised Tsagan-Turutuin-gol valley, but it does not dip towards the mountain margin.

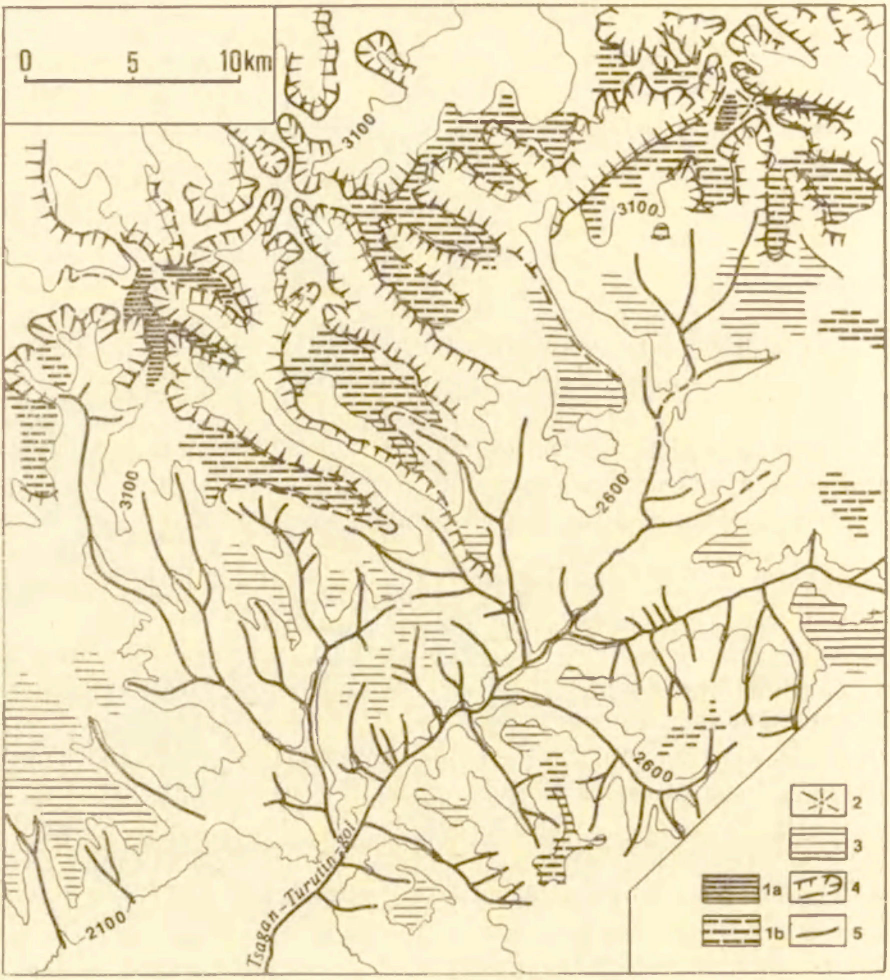

Fig. 4. Major features of relief of the southern Khangai in the Tsagan-Turutuin--gol drainage basin

1 — remnants of planation surfaces, a — upper watershed planation, b — lower watershed planation: 2 — residual hills rising above the upper watershed plateau; 3 — high valley-side flattenings; 4 — upper edges of both glacial cirques and formerly glaciated valleys; 5 — axes of fluvial valleys

On the outer margins of the slightly sloping planation surfaces series of cryoplanation terraces have developed.

Within the internal zone of the Khangai the valley network shape is rectangular. The valleys trend southeast and southwest. For the most part they were modified by glaciers. Glacial cirques, 1—2 km across and 200—450 m deep, lie at 2900—3100 m asl. (Klimek 1977). These may be dammed either by a rock lip or by recessional moraines. Glacial cirques pass into typical glacial troughs, 1.5—2 km wide and 300—500 m deep. The larger glacial troughs may attain lengths of 14—18 km, the smaller

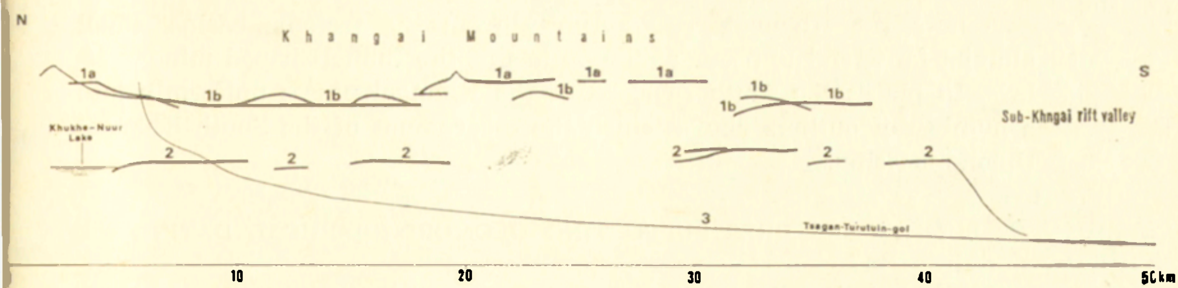

Fig. 5. Planation surfaces developed on the southern slope of the Khangai
1a — upper watershed planation; *1b* — lower watershed planation; 2 — high valley-side flattenings

troughs are 5—7 km long. Along the walls of the troughs lateral moraine ridges are lying at 200—250 m above the floor. Downvalley, at the terminal of the former glaciers, lateral moraines may merge with parallel, crescent-shaped frontal moraines. Recessional moraines occur only in the uppermost parts of the glacial troughs.

The external zone of the Khangai is up to 20 km wide. The initial slope of the Khangai has been deeply eroded into several isolated minor mountain groups, and for this reason steep-sided ridges here are the most important feature (fig. 4). Remnants of the sub-watershed planation surface (3100—3200 m asl.) can be followed for several kilometres on the broad (to 1000 m) ridges. To the south this surface breaks off suddenly on the steep scarp of the Khangai which rises abruptly from the tectonic basins. Within the external zone of the Khangai, the high valley-side flattenings (2600—2800 m asl.) are generally well preserved in the summit areas of the lower mountain groups. The surface discussed does not slope towards the mountain margin. To the west of the Tsagan-Turutuin-gol valley, this surface is wide-spread on the less dissected border of the tectonic Bayan-Nuurin-khotgor Basin. Probably the flattened areas in the South-Khangai Upland, which extends in front of the mountains, can be correlated to the high valley-side flattenings discussed. These are, however, poorly preserved along the Uldzeitu-gol gorge.

Within the external zone the east-west trending valleys have asymmetrical sides (Starkel 1975). North-facing slopes are straight in plan, whilst the steeper opposite slopes are concave in plan with occasional scarps and rock cliffs. The floors of the major valleys at 2100—2200 m asl. are up to 1000 m wide. Around Mandal three rock-cut terraces with a gravel veneer are identified on the Tsagan-Turutuin-gol river at 8—15 m, 20 m and 30 m (Starkel *et al.* 1975). The lowest terrace is very well preserved on the left bank of the stream. This terrace can be traced downstream to the Bayan-Nuurin-khotgor Basin (Klimek 1977). Both alluvial fan deposits and slope deluvia rest on the terrace surface along the scarp foot.

Where the Khangai rises abruptly from the tectonic basins, small alluvial fans tend to occur at the outlets of the funnel-shaped minor valleys. In places the southern slope of the Khangai passes uniformly into a number of outliers above which the escarpment of the South-Khangai Upland is rising.

SUMMIT FEATURES SUPPORTING GEOMORPHOLOGICAL DATING

No correlation deposits are preserved in the Khangai Mts, and for this reason it is not possible to date the surfaces considered. The only evidence are Lower Cretaceous fine-grained deposits with coal seams occurring nearby Khutshirt to the east of the mountains as well as sand and silt series of the same age which survived in the Orkhon graben. These indicate that what is now the Khangai at that time was a levelled surface only slightly rising above the surroundings. According to Soviet workers (Kozhevnikov *et al.* 1970; Selivanov 1972; Khasin 1973) Tertiary tectonic disturbance of block type has been responsible for initiating the mountain massif. Thus the Cretaceous—Lower Tertiary planation surface has been raised. At present its relics are preserved in the summit areas of the Khangai at altitudes of 3000—3500 m asl. These areas have basalt caps of Upper Tertiary age (Kozhevnikov *et al.* 1970). On the northern side of the central Khangai ridge exposures in the glacial cirque walls revealed that the basalt cap is 30—400 m thick and rests on both higher and lower watershed plateaus being separated by a gentle slope. This may suggest low rates of tectonic uplift of the central Khangai ridge during the Upper Tertiary. The most extensive uplift occurred probably during the Quaternary. In Upper Tertiary time the initial summit area of the Khangai was an undulating upland surface with isolated residual hills, having a relative relief of 200—300 m. Such a relief is today typical of the plateaus in eastern Mongolia. It appears that Pleistocene alteration did not obliterate the major features of this inherited relief.

The high valley-side flattenings occur only in the valleys of the Uldzeitu-gol and the Tsagan-Turutuin-gol (fig. 4, 5) which run at right angles to the central Khangai ridge. The position in the landscape suggests that this surface developed along the consequent valleys which drained the Khangai toward the south. It appears that in the southern Khangai there exist two major planation surfaces, and not one surface as suggested by Starkel (1975). The great width of the high valley-side flattenings (fig. 5, 6) and the lack of slope towards the mountain margin may suggest that this surface is in part due to tectonics. It developed in a transverse tectonic depression. Along this initial form consequent valleys were aligned running directly southward. The occurrence of younger tectonics is clearly indicated by the presence within the internal zone of a wide depression which corresponds in height to the flattenings discus-

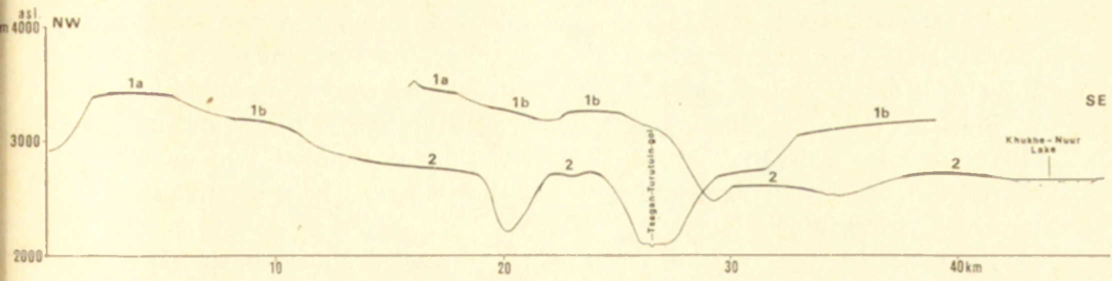

Fig. 6. Sections across the Tsagan-Turutuin-gol drainage basin

1a — upper watershed plateau; *1b* — lower watershed plateau; *2* — high valley-side flattenings

sed. This depression, up to several kilometres long, is occupied by Lake Khukhe-nuur. It is probable that such a distribution of the surface reflects the block structure of the area.

Planation surfaces can be traced from the southern slope of the Khangai Mts to the South-Khangai Upland. The upland surface dips towards the Lake Valley. In this depression one planation surface was identified at 1750 m asl. (Kozhevnikov *et al.* 1970). In the Tsagan-Turutuin-gol drainage basin the planation surfaces break off suddenly on the steep tectonic scarp of the Khangai which rises from the Bayan-Nuurin-khotgor Basin. This indicates that the sub-Khangai rift valleys are due to post-planation surface tectonics. If the lowest surface was in existence in the "Eopleistocene" (this word is Russian and means Upper Tertiary), then the formation of the rift valleys in the mountain foreland was at least early Pleistocene. This conclusion supports the views of Kozhevnikov and others (1970, page 158) that in the Khangai the position of the Continental Divide in early Pleistocene times was different from the present.

QUATERNARY VALLEY FEATURES REFLECTING YOUNG TECTONICS AND CHANGES OF CLIMATE

The formation of the sub-Khangai rift valleys provided greater relief, and the external zone of the mountains was extensively dissected. Valleys have largely destroyed the planation surfaces that have once existed there.

These old surfaces now survive as a series of patches on the Tsagan--Turutuin-gol. Within the internal zone dissection was less effective. Erosion has gone no further than incising the floors of the existing valleys. Remnants of the unrejuvenated old valleys cut slightly into the lowest planation surface (fig. 4) can be found nearby Lake Khukhe-nuur in the headwaters of the Uldzeitu-gol.

In the central part of the Khangai the relief was strongly affected by Pleistocene glaciations. It is now nearly a hundred years since the glacial landforms and deposits in this area were first recognized by Pevtsov (1883) and Potanin (1893), the results of whose work were summarized by Marinov (1954), Marinov and Selivanov (1970). Kozhevnikov and others (1970) and by Selivanov (1972). These writers postulated two or three glaciations of the Khangai and believed that ice caps here occurred during the previous glaciations.

Recent field work in the Tsagan-Turutuin-gol drainage basin (1974—1975) made it possible to determine the nature of the last glaciation in the Khangai (Klimek 1977; Klimek and Sugar 1975). Two generations of end moraines belonging to the Last Glacial are found in the glacial troughs. The older end moraines, 30—70 m high, extend down the valleys to 2200 m asl. Their valley trains pass into the first terrace rising above the actual valley floor. The younger frontal moraines occur only in the major valleys having deep and wide glacial cirques at their heads. Where glacial cirques are shallow, no such moraines exist. The younger moraines comprise 6—9 ridges which attain heights of 10—30 m. The two generations of moraines discussed probably correspond to two stages of the last Khangai glaciation being separated by a longer interval. During this interstadial period the existing glaciers either disappeared completely or survived only in the glacial cirques. Such an interpretation is supported by the lack of younger end moraines in the valleys that have escaped glaciation during the second phase.

At that time new glaciers developed in the valleys where orographic and thermal conditions were favourable for both accumulation and preservation of firn and ice. A characteristic feature is the seldom occurrence of young recessional moraines in the upper parts of the repeatedly glaciated valleys. This may suggest that very dry climatic conditions prevailed during the final phase of the Last Glacial in this part of Asia. Because of moisture defficiency the valley glaciers ceased to transport materials down the valleys and to deposit recessional moraines. Stagnant glaciers and the areal type of deglaciation are also indicated by the frequent occurrence in the glacial troughs of a typical pitted ground moraine.

Glacio-fluvial sediments related to the Last Glacial are found below the oldest terminal moraines within the internal zone and in the Tsagan--Turutuin-gol valley within the external zone. Meltwater streams flowed down this valley carrying sediments on to the mountain foreland. In the middle part of this valley the river passes into a deep gorge, and there is a break in continuity between the outwash fills above the gorge and those below it. Probably the glacio-fluvial terrace which stretches back to the older terminal moraines of the Last Glacial corresponds to the 8—15 m terrace on the Tsagan-Turutuin-gol in the Khangai foreland. This terrace here consists of well rounded gravel and boulders, up to 1 m across. Such coarse glacio-fluvial materials were deposited by very powerful streams. These were proglacial rivers fed by meltwaters during the maximum extent of very active glaciers. Probably, this outwash fill was subsequently dissected by meltwaters which carried a smaller load during the final phase of glaciation. Thus the river was able to erode the earlier outwash fill.

In the Bayan-Nuurin-khotgor Basin the glacio-fluvial terrace discussed above can be traced for several kilometres along the left bank of the Tsagan-Turutuin-gol. This terrace passes into one of the highest terraces of a former lake which once occupied this basin (Klimek and Sugar 1975; Rotnicki 1977). Research in the adjacent Tot-Nuurin-khotgor Basin (Klimek and Rotnicki 1977) revealed that the topmost lacustrine sediments were laid down about 17 000 years ago ($^{14}C = 17\,220 \pm 155$ years B.P.Hv).

The post-glacial period (Holocene) was one of lesser alterations of the Khangai valleys. At higher levels, where the glacial cirques and troughs occur valley floor changes were minimal. At lower levels with predominant fluvial features the older outwash fills were dissected. In places channel reaches are also incised in bedrock.

In the major valley floors, especially in the Tsagan-Turutuin-gol valley floor, traces of dry, now abandoned channels are observed. These indicate that in former times the stream discharges were far greater than they are now. The present streams tend to adjust themselves to the existing channel features. It is likely, that in the immediate past the valley floors were refashioned by catastrophic floods with a recurrence interval of tens of years. However, high discharges can also be related to a more humid phase of the Holocene during which the present valley floor relief developed. The valley bottom is strewn with blocks and coarse gravels washed out from the valley fills. In former times this debris was considered to be the youngest outwash deposit (Klimek and Sugar 1975).

SLOPE FEATURES AS INDICATORS OF CLIMATE CHANGES

In the southern Khangai there occur slopes of different ages and of different origins. Original slopes have been refashioned by various present-day and past morphogenetic processes to a different degree.

Within the internal zone rising above 2700 m asl. long slopes with gradients of 10—12° separate the higher watershed plateau from the lower sub-watershed plateau. This distinctive landform assemblage is the oldest feature in the southern Khangai. Under cold climatic conditions these slopes have been heavily modified by periglacial processes. Wherever lithologic conditions were suitable series of cryoplanation terraces developed on the slopes. At the present time frost action processes and congelifluction continue to modify such slopes (comp. chapter VI).

It appears, however, that the intensity of actual processes is minor than during the Last Glacial. The internal zone comprises both slopes of Tertiary foundation and very young slopes enclosing the glacial cirques and troughs. The latter slopes resulted greatly from the modification of the older valley-sides by valley glaciers during the Last Glacial. Such slopes are steep, attaining heights of 300—500 m. In places rock cliffs, and even polished surfaces occur. At present slopes are being altered by intense rock and waste fall. This process is controlled by slope aspect and jointing (on granite). The lower slope sectors and footslopes are covered with fallen blocks which tend to build up the lateral moraines along the sides of glacial troughs.

The flanks of end moraines and of some lateral moraines which consist of rock debris and clay are incised by numerous channels. These are drained by ephemeral streamlets. It appears that the channels were cut during a former phase of higher rainfall intensities.

Within the external zone (below 2700 m) the dominant slopes are stream valley-sides. Slope evolution here was influenced by changes of climate which took place during the Quaternary. Steep slopes occur commonly. They are notched by ravins of corrasional type.

On slopes that have southerly exposures and a mountain steppe vegetation either blocky or granular waste covers occur. North-facing slopes have patches of larch forest and a clayey-scree waste cover. Both lithology and slope aspect control nature and thickness of scree sheets to a large degree. These slope aspect contrasts (asymmetry) may be so strong as to produce quite different slope sheets in the east—west running valleys (Starkel 1975; Kowalkowski *et al.* 1977). Under cool conditions of the Pleistocene the efficacy of slope processes was greater on slopes facing the sun that have been underlain by permafrost. This is indicated by the presence on such slopes of both fossil congelifluction lobes and rock falls (Kowalkowski *et al.* 1977), and of inactive cryoplanation terraces with a blocky-granular overburden (Pękala 1979). The clayey-scree sheets gave rise to parachernozems (Kowalkowski and Lomborinchen 1975). The north-facing slopes underlain by permafrost have only a fine-grained overburden being affected by creep.

The southern face of the Khangai shows steep slopes of tectonic origin. Below the tectonic scarp, in the neighbourhood of the Tsagan-Turutuin--gol valley, where debris was efficiently evacuated cryopediments developed on granitic rocks (2100 m asl.). These cryopediments pass into the youngest Pleistocene terrace. From this one may postulate a cooler and moister climate at the time of cryopediment formation.

Under the cooler and more humid conditions of the Pleistocene both formation and transformation of the mountain slopes took place within two different morphoclimatic belts. At this time, the snow line deduced from lowest cirque-floor levels (Klimaszewski 1973) has been at about

2800—2900 m asl. Since the summit plateaus reached to much higher altitudes, they must have been covered with permanent snow patches, and here valley-type glaciers appeared. From the larger cirques glaciers descended to much lower altitudes (2400—2500 m asl.) causing valley-side modification. The presence of both fossil cryoplanation terraces at 2400 m asl. (Pękala 1975) and fossil cryopediments at 2100 m asl., which tend to form under similar conditions, implies a lower position of the periglacial belt. Its lower limit fell to 2100 m asl. It seems that at that time the whole Bayan-Nuurin-khotgor Basin has also experienced periglacial conditions.

The post-glacial climatic warming has caused a rise of the snow line of a least 600—700 m. The glaciers disappeared and the periglacial limits moved upslope at the same rate. At present less intense frost action and congelifluction processes as well as active cryoplanation terraces occur only above altitude of 2700 m asl. (comp. chapter VI).

During the Holocene the rate of slope transformation is correspondingly less. A more humid phase which occurred between 5000 and 2000 yrs B.P. (Vipper *et al.* 1976) within the northern piedmont zone may be expected to have had some effect upon the southern Khangai as well. At higher levels this wetter phase was marked by increased development of ravines of corrosional type on the moraine ridges and bedrock. The subsequent climatic cooling about 2000 yrs B.P. was accompanied by increased aridity. This resulted in both small-scale frost weathering and formation of fine-grained talus below the rock cliffs that occur within the external zone of the Khangai (Kowalkowski *et al.* 1977).

ELIGIUSZ BRZEŹNIAK, TADEUSZ NIEDŹWIEDŹ

IV. VERTICAL VARIABILITY OF CLIMATIC CONDITIONS IN THE KHANGAI MOUNTAINS

The Khangai Mountains are situated in the southern Khangai—Khubsugul climatic region which experiences moderate continentality and cool and very cool winters (Badarch 1971; Gungaadash 1971). January mean temperatures vary here from −15° to −25°, the mean diurnal temperatures in July are not above 15°C, and the annual mean temperatures are below −4°C. Precipitation amounts range from 250 to 400 mm per year. The Khangai rises from 2000 m to 3540 m asl., and for this reason the climatic phenomena are arranged belt-like.

By using the classification of Koppen and Geiger (*vide* Blüthgen 1966) the southern Khangai foreland is classified as having a BS dry steppe climate with cool winter. The mountains rising up to 2700 m asl. have a DW — cold snowy forest climate with dry winter; annual mean temperatures here are below −3°C, and July mean temperatures are above 9—10°C. The mountain slopes and tops which extend above 2700 m asl. experience an ET — tundra climate. Such conditions are typical of continental areas where the mean temperature of the warmest month is less than 9°C (*ibidem*).

In Mongolia, the continentality of climate is reflected principally in the wide variations of both annual and diurnal air temperatures. According to Ivanov's continentality index[1] (*vide* Avirmid 1970), the interior of the Khangai is determined as a moderately continental region (value below 205). The southern slope of the Khangai belongs to the very continental regions (values between 205 and 250), while the southern Khangai foreland is defined as a highly continental region (values above 250).

The values of Budyko's radiational index of dryness[2] (1971) range

[1] The index of continentality has a value of 100 when the continental influence is in equilibrium with the oceanic influences.

[2] Budyko's radiational index of dryness (1971) is expressed by the ratio of the net radiation available to evaporate water vapour from a wet surface to the heat required to evaporate the mean annual precipitation.

from 1.5 at Tsetserleg to 2.1 at Khuzhirt (steppe zone) on the northern slope of the Khangai and to 2.5 at Galuut on the southern slope. The latter index is greater than the value for the semi-deserts because of the high values of the net radiation balance.

Summer temperatures are largely dependent on altitude and latitude. The relationship between the July mean temperature and the height above sea level (H), latitude (φ) and longitude (λ) for 30 stations can be expressed by

$$t_{VII} = 62.2 - 0.0070\,H - 0.648\varphi - 0.042\lambda$$

The calculated correlation coefficient being $r = 0.976$ and the estimation standard error $B_{es.} - 1.1$.

The mean July temperature variations in the Khangai are shown in figure 2. It appears that this temperature decreases with height at the rate of 0.7°C/100 m and poleward at the rate of 0.6°C/1° of latitude.

The chapter deals with the height-controlled variations of temperatures, with local air streams and with atmospheric phenomena occurring on the southern slope of the Khangai. Discussion of microclimatic problems will be based on data obtained from the Galuut and Tsetserleg stations (*Climatic Annual 1971*).

Mesoclimatic data for the periods: 21 June—31 July, 1974, and 1 July—15 August, 1975, were collected at four sites situated at altitudes of 1950—3352m (tab. 1; fig. 7). The basic Mandal station was located on the Tsagan-Turutuin-gol river where it emerges from the mountains, at 29 above the valley-floor.

Table 1. Location of observation spots

Observation spot	Altitude (H)	Latitude (φ)	Longitude (λ)	Vegetation	Period of record
Donoin-Dzun--Nuruu	3352 m	47°05′	100°12′	Stony mountain tundra, plant cover ca. 5%	1 July — 15 August, 1975
Olon-nuur	2650 m	47°04′	100°15′	Mountain steppe, plant cover ca. 80%	1 July — 15 August, 1975
Galuut	2117 m	46°42′	100°08′	*Stipa Krylovii—Artemisia Frigida* steppe, plant cover ca. 80%	21 June — 31 July, 1974 1 July — 15 August, 1975
Mandal	2055 m	46°50′	100°05′	*Stipa Krylovii—Artemisia Frigida* steppe, plant cover ca. 80%	21 June — 31 July, 1975 1 July — 15 August, 1975
"Basin"	1950 m	46°39′	99°55′	Steppe, plant cover ca. 75%	21 June — 31 July, 1974 1 July — 15 August, 1975

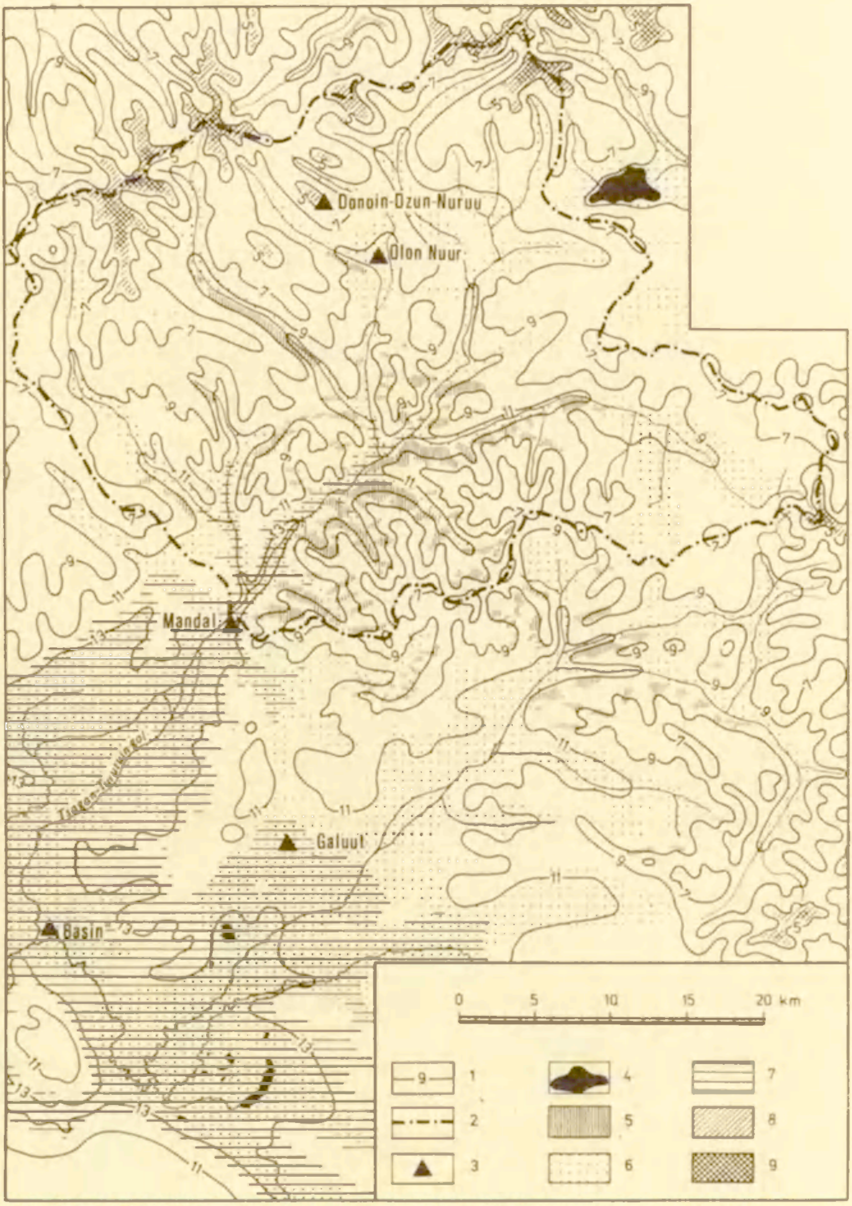

Fig. 7. The distribution pattern of thermal belts on the southern slope of the Khangai in July

1 — isotherme of the mean July air temperature corresponding to boundaries of the thermal belts; 2 — the Tsagan-Turutuin-gol basin divide above Mandal; 3 — observation spot; 4 — lake; 5 — larch forest; 6 — area with strong temperature inversions; 7 — area where maximum air temperatures in July are always above 10°C, > 20% of days have $t_{max} > 20°C$, and more than 1% of days have $t_{max} > 25°C$; 8 — area where $t_{min} < 0°C$ occurs on more than 50% of days in July; 9 — area where maximum air temperature may drop below 0°C in July

VARIATIONS OF SELECTED CLIMATIC ELEMENTS

CIRCULATION

The atmospheric circulation over Mongolia depends on a number of factors including position in the continental interior, great distance from the oceans and the existence of mountain barriers.

In winter, Mongolia is dominated by the large Asiatic anticyclone leading to very low air temperatures (to −50°C), low cloudiness and a total lack of precipitation. Sometimes warmer air masses coming from China may disturb this winter high. Local cyclones causing snow fall, strong winds and snow storms tend to develop then over regions where the two contrasting air masses meet, especially in the mountains (Zhadambaa *et al.* 1967).

In the spring, i.e. at the beginning of April, the weather in Mongolia is affected by the temperate latitude Asiatic front moving from the south poleward. At this time a high of the polar continental type occurs over eastern Siberia. Steep temperature gradients existing between these two pressure systems lead to the formation of both waves and local lows along the Asiatic front. There also occur cold fronts which are linked with the lows moving in the temperate latitudes over Siberia. In spring the complex synoptic processes are responsible for the great variability of weather including cloud cover, temperature, pressure and increased wind velocities being accompanied by dust storms.

Summer is characterized by the increased activity of the southerly and southwesterly lows which often are associated with frontal zones. These are slowing down over the mountainous regions, and for this reason conditions are favourable for the formation of wave patterns. Marked activity of the Asiatic front with cumulonimbus clouds, abundant rainfall and thunderstorms occurs during this season over Mongolia. In the Khangai, summer rainfall also is due to the intense convection of heat on the mountain-sides.

In autumn, as the high develops, the Asiatic front advances south and southeast. Towards the middle of October the winter anticyclonal weather regime tends to extend across Mongolia.

WINDS

In the Khangai, the general direction of wind travel does not correspond to that of the low-level winds. The movement of the lowest layer of the air currents is largely modified by the surface configuration and type of the ground surface. This interrelationship is illustrated by the annual mean frequency of wind occurrence at Galuut on the southern Khangai slope and at Tsetserleg on the northern slope of the Khangai (*Climatic Annual 1971*). At Galuut there prevail northern (14.6%) and northwestern (12.6%) winds, while southeastern winds are inconsiderable (1.1%). At Tsetserleg northwestern (30.4%) and western (11.3%) winds are dominant, whereas northeastern (1.3%) and southern (1.4%) winds less frequent. At both stations calms occur frequently: at Galuut 53.5% and at Tsetserleg 31.6%.

The varied relief of the southern slope of the Khangai, and especially the arrangement of valleys greatly affect the movement of air currents in the warm season. Consequently, northern, northeastern and southwestern winds prevail at the Galuut, Mandal and "Basin" stations. This influence of local topography on the prevailing wind direction was stated in both research seasons of 1974 and 1975 (Avirmid *et al.* 1975, 1976). Eastern and southeastern winds were least frequent there. At Olon-nuur northern, southern and southwestern winds were dominant, whereas southeastern winds occurred occasionally. A typical feature of the movement of air is the variable frequency of calm occurrence within the piedmont basins and valleys (at Galuut 55—70% and at "Basin" 15—19%). This means that in the basins including the Galuut and "Basin" stations the drainage of air is rather impeded, whereas the Tsagan-Turutuin-gol valley is freely ventilated (only 2% of calms in 1974 and 0.5% in 1975). At Olon-nuur the observed percentage value of calm occurrence was 20 in July, 1975.

Over much of the area there is a mountain-valley circulation regime. At both Galuut and Mandal diurnal variations of the air currents (i.e. the clockwise reversal of wind direction) were observed. During the day upvalley, southwestern winds blew causing the convectional clouds to develop over the mountain ridges. At night radiation caused cooling of the air which moved downvalley towards the sub-Khangai basins.

RADIATION

The annual total of sunshine received by the Khangai is 2680—2731 hours (data for Khuzhirt and Tsetserleg; *Climatic Annual 1971*). The monthly sunshine values show little change throughout the year and vary slightly from 225 to 300 hours during the months March—October. For the remaining part of the year the monthly totals are no less than 100—150 hours, since Mongolia is dominated by the anticyclone.

At Galuut, on a horizontal surface, the global radiation value is about 136 kcal/cm^2 per year (Badarch 1972), and in the surroundings of Tsetserleg the corresponding value is 124 kcal/cm^2 per year. The global radiation varies with the seasons from 4,2 kcal/cm^2 in December to 18.2 kcal/cm^2 in June. The net radiation reaches values of 40—41 kcal/cm^2 on the southern fringe of the Khangai and of 30—38 kcal/cm^2 on its northern fringe. Values of the radiation balance are negative from September until February. The highest values of 8.5 kcal/cm^2 are found in July. In

July, 1974 the radiation balance at Mandal reached a value of 7.6 kcal/ /cm², and in July, 1975, it was 9.6 kcal/cm². The corresponding values of the albedo of the steppe surface were 23 and 21%. Measurements made in July, 1974, revealed that during the daytime about two-third of the radiation balance is spent in heating the air through turbulent heat exchange, and about one-third of the radiation balance is available to evaporation from the ground surface. About 0.09 cal/cm². min are used for heating the soil.

THERMAL CONDITIONS

Both annual averages from Galuut and Tsetserleg and data obtained by mesurement of temperature distribution and its annual changes on the southern mountain slope in 1974 (21 June—31 July) and in 1975 (1 July—15 August) give some impression of the temperature variations in the Khangai. The results obtained from different stations correlate with those at Galuut, and for this reason it was possible to reduce the July temperatures recorded in 1974 and 1975 in relation to the observation period of 1957—1975 (Brzeźniak 1977).

The varied relief has a marked influence upon air temperature in the Khangai and in the piedmont basins. Consequently, the annual mean temperature is −5.3°C at Galuut in the southern foreland of the mountains and only −0.1°C at Tsetserleg on the northern mountain slope. After some modification to allow for the height difference of 420 m between both stations it appears that the annual mean temperature at Tsetserleg is some 2°C higher than that at Galuut.

Throughout the year the temperature differences between both stations vary from 10.5°C in January to 1.8°C in July. The annual average maximum temperatures are 3.4°C at Galuut and 6.8°C at Tsetserleg. The highest temperatures (19.3°C at Galuut and 21.2°C at Tsetserleg) are reached in July, while negative values are observed from November until March. The annual mean minimum temperature at Galuut is by 6°C lower than that at Tsetserleg (−6.1°C) because of the cold air stagnation in the Galuut Basin. The lowest mean minimum temperatures are recorded in January (−32.2°C at Galuut and −21.4°C at Tsetserleg). Positive values occur only during the months June—August at Galuut and May—August at Tsetserleg. Absolute maximum temperatures of 32°C (at Galuut) and of 34°C (at Tsetserleg) are recorded in July, and absolute minimum temperatures of −48°C and −38°C are observed in December.

The work of Brzeźniak (1977) has confirmed the belt-like arrangement of components of the temperature régime on the southern slope of the Khangai. Data calculated for 46 days in 1975 indicate that the daily mean air temperature varied from 13.8°C (at "Basin") to 13.4°C (at Mandal) and to 4.4°C on the mountain tops, i.e. on the Donoin-Dzun-Nuruu ridge.

The temperature diminished with height at the rate of 0.7°C/100 m. The same applies to the distribution of temperatures in the summer of 1974. The daily mean temperature was then 14.4°C at Mandal and 14.6°C at Galuut. The maximum, minimum and daily air temperature ranges also decreased with height. In 1975 the average maximum air temperature varied between 19.9°C (at "Basin") and 8.3°C (on the Donoin-Dzun-Nuruu). The average vertical temperature gradient was 0.5°C/100 m. The highest average minimum temperature of 7.0°C was recorded at Mandal, while the lowest value of 1.1°C was found on the Donoin-Dzun-Nuruu ridge. The diurnal mean air temperature ranges decreased with height from 14.4°C at "Basin" to 7.2°C on the Khangai mountain tops. In the research period the absolute maximum temperature of 27.9°C was observed at Mandal, while the absolute minimum temperature of −7.0°C occurred on the Donoin-Dzun-Nuruu.

Throughout the 24 hour cycle the highest temperature was recorded at 16 h and the lowest temperatures at 6 h on the whole southern slope of the Khangai. Temperature differences decreased with height from 12.3 (at "Basin") to 9.2 (at Olon-nuur) and to 5.2°C (on the Donoin-Dzun-Nuruu).

The extreme continentality of climate and the presence of a stationary high in winter produces strong temperature inversions in Mongolia. In Ulan Bator inversions occurred during 94% of all days in January, 1958 (Makhover 1967). Temperature differences were on average 14°C in the lowest 960 m layer of air. These temperature inversions occur most frequently in the lowest 250 m layer of the atmosphere, where the temperature differences may exceed 10°C on 73% of all days in January. In the inversional layer temperature differences may reach as much as 32°C (Zhadambaa 1972). In the Khangai, January mean air temperatures of −15°C are likely to occur at altitudes of 2500—2600 m. The above discussed type of air temperature distribution in the mountainous regions of Mongolia persists throughout the year. Because of strong winter inversions which have become known as a singularity of the mesoclimate of Mongolia (Gavrilova 1974) the Khangai tends to be some 10°C warmer than the surrounding areas. It is likely that even the highest mountain tops are some 5°C warmer than the sub-Khangai basins.

Measurements also revealed the occurrence of temperature inversions in the summer season. For instance, on June 24, 1974, during a cloudless night at 23 h a temperature of 9.7°C was recorded in the Tsagan-Turutuin-gol valley-bottom, while 250—300 m further upslope the temperature was 2.7°C higher than that in the valley-bottom.

The inversional distribution of minimum temperatures was detected along the "Basin"—Mandal profile with a height difference of 105 m (fig. 8). The inversions were associated principally with a slightly (in 0—30%) clouded sky with low-level clouds. Over the period 1 July-15

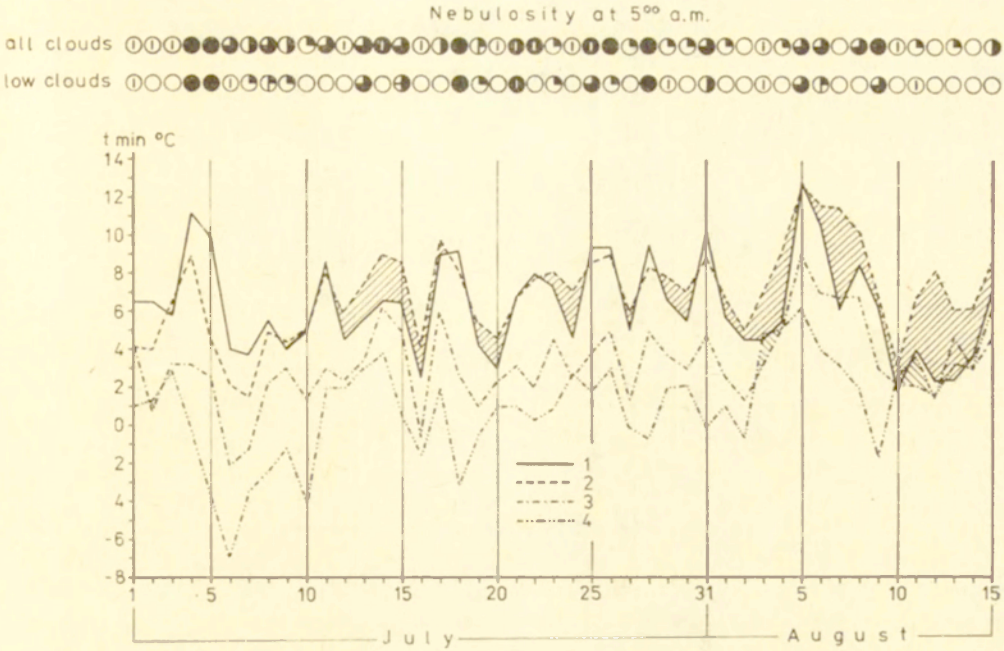

Fig. 8. Diurnal minimum air temperature: at "Basin" (*1*), Mandal (*2*), Olon-nuur (*3*) and on the Donoin-Dzun-Nuruu ridge (*4*); 1 July—15 August, 1975

August, 1975, minimum temperature inversions existed on 65% of days. Slight inversions of the order of 1.0—1.9°C were dominant. Most distinct (5.6°C) was the inversion on August. The upper parts of the southern Khangai experienced inversions only five times when the minimum temperatures on the Donoin-Dzun-Nuruu ridge were higher than those in the Olon-nuur valley, a height difference of 702 m.

Throughout the 24 hour cycle temperature inversions began to develop as early as 20 h (2.8% of inversion conditions in the "Basin"—Mandal profile). As the cold air lake in the piedmont basin developed, the inversion became more intensive. In general, inversions occurred most frequently early in the morning at 5—7 h (10.7—12.5%) when the temperature differences showed a maximum of up to 7°C. Very slight inversions (up to 0.9°C) and slight inversions (1.0—1.9°C) were dominant (tab. 2).

In the summer season temperature inversions also occur on the northern slope of the Khangai (Beresneva 1977a). In minor depressions the minimum temperatures were on average by 6°C lower than those on the surrounding convex landforms.

The southern fringe of the Khangai experiences first frost at the end of August or the beginning of September, while the last frost is recorded in the first decades of June. The frost-less period in the basins lasts on average for only 64 days (at Galuut). In the northern Khangai foreland

Table 2. Frequency of occurrence in the Khangai Mts of temperature inversions between the stations "Basin" and Mandal (1 July—15 August, 1975)

Intensity [°C]	Hours													Total	Percentage
	20	21	22	23	24	01	02	03	04	05	06	07	08		
0.0—0.9	5	8	8	8	13	13	14	9	7	9	14	9	15	132	45.6
1.0—1.9	1	2	—	2	4	12	8	10	10	10	8	11	8	86	29.8
2.0—2.9	2	—	1	2	1	1	3	1	4	8	6	7	3	39	13.5
3.0—3.9	—	—	—	—	1	1	1	3	3	1	4	2	—	16	5.5
4.0—4.9	—	—	—	—	—	1	1	—	—	2	2	1	1	8	2.7
5.0—5.9	—	—	—	—	—	—	—	2	2	1	1	—	—	6	2.1
6.0—6.9	—	—	—	—	—	—	—	—	—	—	—	1	—	1	0.4
7.0—7.9	—	—	—	—	—	—	—	—	—	—	1	—	—	1	0.4
Total	8	10	9	12	19	28	27	25	26	31	36	31	27	289	
Percentage	2.8	3.5	3.1	4.2	6.6	9.7	9.3	8.6	9.0	10.7	12.5	10.7	9.3	100.0	100.0

the frost-less period lasts for about 102—109 days. On the convex landforms it lasts about 50 days longer than in the valleys and basins (Beresneva 1977b). On the main Khangai ridge with rises above 3300 m asl. temperature may fall below 0°C during the whole day, even in July, while above 3400 m asl. frost occurs on 50% of days out of 31 days in July.

A characteristic feature of the area discussed is the marked length of the thermal winter. It begins on average about 1 October and lasts for 201 days until 24 April at Galuut, but at Tsetserleg the winter begins about 18 October and lasts for 171 days until 8 April. The growing season with temperatures above 5°C starts about 2 May in the Tsetserleg region and towards the middle of May at Galuut, and lasts for between 149 and 124 days until 15—27 September. In the interior of the Khangai at about 3250 m asl. the growing season is not present. The period when temperatures rise above 10°C lasts for between 73 and 98 days in the Khangai foreland, and the totals of temperatures for this period are 850—1300°C, diminishing to zero at altitudes of about 2500 m with any thermal summer at all.

TEMPERATURES OF THE GROUND SURFACE AND SOIL TEMPERATURES

Slightly clouded skies are suitable conditions for marked differences in both heating and cooling of the ground surface. The diurnal temperature variations of the ground surface can be above 60°C. In July, 1974, the very intense direct solar radiation caused the temperatures of the dry steppe surface to rise as high as 60°C, whereas at night temperatures fell as low as 0.4°C. In 1974, the average diurnal temperature ranges of this surface reached 38°C with a maximum of 43.4°C and a minimum of 5.5°C, diminishing to 3.3°C at 20 cm soil depth. The lag time of the soil temperature maximum was six hours. Data obtained for July, 1970—1975, indicate that at Galuut temperature variations of the ground surface were above 50°C on 10% of all days falling below 26.2°C on 10% of days. Minima were below −0.7°C and above 8.9°C on 10% of days. Diurnal temperature ranges were above 46.9°C on 10% of days. In the higher part of the Khangai the July maximum soil temperatures at the Olon-nuur station (2650 m asl.) were about 8°C lower than those at Galuut. Similarly, the minimum soil temperatures were reduced by 0.7°C. The average gradient between the "Basin" and Olon-nuur stations was −0.74°C/100 m. Data obtained by measurements of the soil temperatures in 1975 show that the average values of the gradient between the Mandal and Olon-nuur stations ranged from −0.9°C/100 m at 50 cm soil depth to −1.03°C/100 m at 10 cm soil depth. At the same time soil temperature differences at 5—50 cm soil depths diminished with height from 8 at "Basin" to 4.1 at Mandal, and to 2.9°C at Olon-nuur.

AIR HUMIDITY

Very low mean annual values of relative humidity (57%) are found at Galuut and at Tsetserleg. The variations of monthly mean values of relative humidity also are unimportant. The humidity differences between Galuut and Tsetserleg are of the order of 1—10% during the months from March to October. In winter the values of relative humidity at Galuut are about 8—11% higher than those at Tsetserleg. The lowest values (i.e. the driest air) were recorded at Galuut in May (38%) and at Tsetserleg in April (47%). Both the annual and monthly mean values of the saturation deficit also vary only slightly. The differences in average annual values between Galuut (3.4 mb) and Tsetserleg (3.8 mb) is 0.4 mb.

The highest monthly values of the saturation deficit were recorded at both stations in June (at Galuut 7.8 mb, at Tsetserleg 7.3 mb), while the lowest values occurred in December and January (0.3—0.4 mb at Galuut and 1.0 mb at Tsetserleg).

CLOUDINESS AND PRECIPITATION

Over the Khangai distinct variations in cloudiness occur with the changing season. Cloudiness maxima are recorded in summer with only 4 clear days in July, while minima with 18 clear days prevail in January.

Observations made at Mandal in the summer seasons of 1974 and 1975 revealed that a distinct daily maximum of cloudiness (72%) exists about 17 h due to the increased thermal convection on the mountain slopes. The lowest cloud amounts are found at 2 h (45%). In the interior of the mountains, during the daytime, cloud amounts are increasing from 69% at Galuut to 81% at Olon-nuur, the average daily cloudiness being 66% and 72%.

From October until March the monthly precipitation totals do not rise above 10 mm, and 67% of the annual amount fall during the three summer months from June to August. The southern fringe of the Khangai receives 240—280 m of precipitation per year, and the interior of the mountains receives more than 400 mm. At Galuut the differences in annual precipitation totals are up to 283 mm. In July, with an average of 78 mm, rain amounts varied between 29 mm and 134 mm. In July, the northeastern slopes of the Khangai receive on average 100 mm of rainfall or above. From May until October, the falls last for about 60 hours in the Khangai (Kadyrova 1976). When compared with Galuut, the summer rainfall at Mandal varied between 154% (in 1975) and 22% (in 1974). In the interior of the mountains it was 162% in the Olon-nuur valley and 208% on the Donoin-Dzun-Nuruu ridge (in 1975). In the summer showers with thunderstorms tend to prevail, especially in the mountains. Over the period 16 June—31 July, 1974, 13 days with thunderstorms were recorded at Mandal and 7 days at Galuut, of which two days were hail and on two days rainfall was above 10 mm.

In the Khangai the snow cover lasts for about 7 months, from September until April, its average depth being only 7 cm in March (BNMAU 1976). Galuut has 177 days with snow lying. In the higher parts of the mountains, the average depth of the snow cover is above 15—20 cm (Badarch 1971) increasing locally to 1 m. On the south-facing slopes sublimation tends to destroy rapidly the snow (Froehlich and Slupik 1977). Above 3300 m asl. snow patches persist throughout the year.

THE VERTICAL DISTRIBUTION PATTERN OF THERMAL PHENOMENA IN THE KHANGAI

A review of the most important elements of climate shows that a number of them undergoes marked changes with height. The most sensitive and height-dependent element is the summer air temperature. In winter the temperature variations are modified clearly by inversions, and for this reason the July temperature forms the base of the following division. On the southern Khangai slope the upper tree line (at 2700 m als.) corresponds to the July isotherm of 9°C. Further limiting values for air temperatures at 2°C intervals are calculated by using the classification method of Hess (1965). The thermal belts on the southern slope of the Khangai (fig. 8, 9) are as follows:

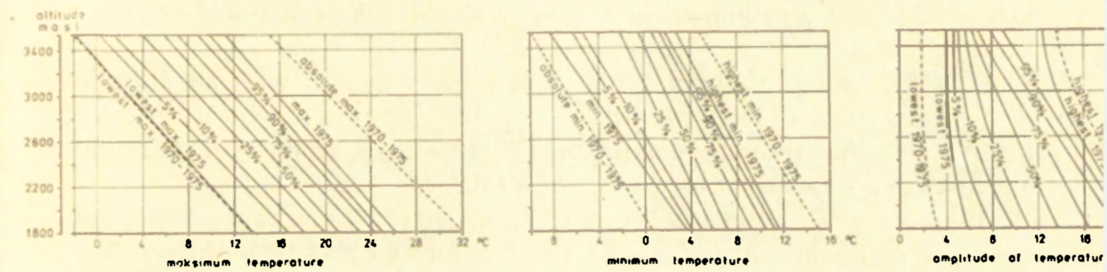

Fig. 9. The probability of occurrence on the southern slope of the Khangai of both maximum and minimum air temperatures and of diurnal air temperature ranges dropping below certain values (based on data for July, 1975). Broken line indicates extreme July values recorded at Galuut, 1970—1975

A — The steppe belt reaches up to 2100 m asl. It has a July mean temperature of above 13°C. The absolute maximum temperature exceeds 29°C. Temperatures rise above 20°C on more than 30% of all days in July, and the absolute minimum temperatures can fall as low as —1.5°C. Frost is likely to occur on 2% of days, and on 25% of days the diurnal air temperature ranges are about 15°.

B — The lower forest-steppe belt approximates with an altitude of between 2100 m and 2400 m asl. and has July mean temperatures of 11—13°C and absolute maximum temperatures of 26—29°C. On 5—30% of all days in July temperatures rise above 20°C, and 1% of days has temperatures above 25°C. Minimum temperatures can fall as low as —3°C, and negative values are recorded on 2—4% of days. Diurnal air temperature variations of more than 15°C occur on 11—25% of days. This belt includes mountain steppe and larch forest communities on north-facing slopes (comp. chapter VIII).

C — The upper forest-steppe belt extends between 2400 m and 2700 m asl. The July mean temperature varies from 9 to 11°C, and the highest temperatures reach 23.5—26°C. On 2—5% of all days in July temperatures rise above 20°C. Temperatures exceeding 25°C occur occasionally in the lower part of this belt. At its upper limit minimum temperatures can be —5°C, and frost is recorded on 4—10% of all days in July. Minimum temperatures rise above 10°C on 1—2% of days. Diurnal air temperature variations of more than 15°C (2—11%) are still recorded there. On the north-facing slopes larch forest patches are found up to altitudes of 2700 m.

D — The high-mountain meadow belt extends between 2700 m and 3000 m asl. and has July mean temperatures of 7°—9°C, and 21—23.5°C are the absolute maximum values. Temperatures rise above 20°C on 1—2% of all days in July, and maximum temperatures of up 10°C are recorded on 10—34% of days. At the upper limit of this belt minimum temperatures can drop to −6.5°C, and frost is recorded on 10—20% of all days in July. Minimum temperatures rising above 10°C are absent. Diurnal air temperature ranges above 10°C are found on 32—52% of days.

E — The stony mountain tundra belt approximates with an altitude of between 3000 m and 3300 m asl. The July mean temperature is here 5—7°C and the absolute maxima reach 18—21°C. Temperatures rising above 15°C are recorded on 2—5% of all days in July.

Temperatures falling below 5°C occur on 3—15% days. Minimum temperatures can drop to −8.5°C. Frost occurs on 20—35% of days. Minimum temperatures exceed 5°C on 2—6% of all days in July. Diurnal air temperature ranges exceeding 10°C are recorded on 9—32% of days. In this belt the vegetation cover is poorly developed and patterned ground occurs in flat sites.

F — The scree belt extends above 3300 m asl. On the ridges the July mean temperature drops below 5°C, being 2.8°C at 3540 m asl. The highest temperatures are 15—18°C, and temperatures below 10°C are dominant (77—95% of all days in July). Temperatures falling below 0°C throughout the day can occur even in July. Frost is recorded on 35—60% of days. Minimum temperatures rising above 5°C are found on 1—2% of days. Diurnal temperature ranges are small and variations of more than 10°C occur on 2—9% of days. Snow patches persist all year round.

The vertical July temperature distribution patterns on the southern slope of the Khangai tend to reflect the major features of the plant cover (Pacyna 1976), of the hydrological phenomena (Dauksza and Soja 1977), of the soil cover (see chapter VII) and of the course of morphogenetic processes (see chapter VI).

ROMAN SOJA

V. AREAL VARIABILITY OF HYDROLOGICAL CONDITIONS IN THE TSAGAN-TURUTUIN-GOL DRAINAGE BASIN, SOUTHERN KHANGAI MOUNTAINS

The Tsagan-Turutuin-gol catchment is contained within the head-waters of the river Baidarak which drains the southern Khangai. The rivers here are characterized by seasonal contrasts in discharges. Low flows prevail during the months September—May, while high flows pre-dominate during the remaining year, with peak flows influenced by rainfall occurrence (Kuznetsov 1959). Comparative hydrographs for two streams draining the southern slope of the Khangai are shown in figure 10. The annual variation in runoff from the Tsagan-Turutuin-gol catchments is likely to be transitional between the two streams. The Tsagan-Turutuin-gol (Kuznetsov 1968) receives its water principally from rainfall concentrated in the summer season (50—60% of the annual runoff value), from ground water supply (15—20%) and from snowmelt (15—20%). W. Froehlich and others (1975) concluded that both springs and icings are clearly important for water supply. It seems, however, that springs and icings have particular significance at the lowest stages, but they are unimportant throughout the year.

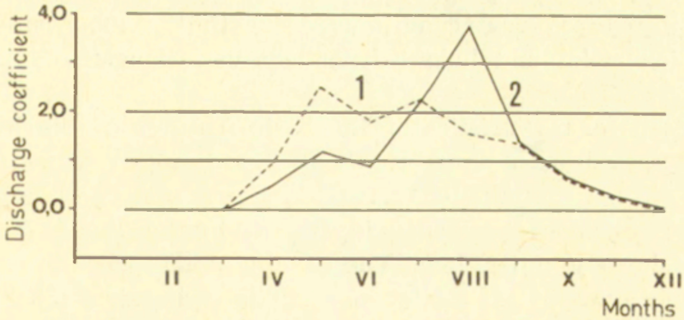

Fig. 10. Mean monthly discharge coefficients for two rivers draining the southern Khangai (based on data by BNMAU 1975)

1 — Ongin-gol; 2 — Tchingestein-gol

In the Tsagan-Turutuin-gol catchment the hydrological cycle is greatly controlled by the marked precipitation deficiency. In the southern mountain foreland, precipitation amounts are 200—300 mm per year (Baatar and Rossomakhin 1968). On the mountain ridges average annual precipitation can be 500 mm in the wetter years. The precipitation gradient is 5 mm per each 100 m increase in height (*ibidem*). The area experiences a summer rainfall maximum (more than 60% annual total), and the precipitation gradient at 2000—3350 m asl. is slightly less than 10 mm/100 m (Dauksza and Soja 1977). In the highest summit areas snowfall is recorded even in July (Brzeźniak 1977). The evaporation and runoff values for the southern Khangai catchments have been calculated by Soviet workers (Kuznetsov 1959, 1968; *The global water balance*... 1974) It is assumed that potential evaporation in the southern foreland of the Khangai exceeds 500 mm, but on the central Khangai ridge it is about 200 mm corresponding to evaporation.

In the Khangai runoff values vary from 50 to 100 mm (Kuznetsov 1959) dependent on the precipitation amount in the given year. There is a general southerly decrease in runoff. In the mountain foreland runoff is less than 200 mm, the runoff coefficient is 0.1—0.2 (*The global water balance*... 1974). These data indicate that to the south all of the hydrological parameters are clearly changing. The individual values may be open to doubt, but the hydrological cycles appear to vary with altitude increasing in the mountains and with aridity increasing southward. Thus it was possible to distinguish two hydrological regions in the Khangai (Kuznetsov 1968): a region of perennial runoff production and a region of decreasing runoff.

It is difficult to draw a line of demarcation between the two regions. Does this boundary correspond to the clear geomorphological boundary between the mountains and the sub-Khangai basins or has the region from which perennial runoff is produced moved northwards? Another problem arises from the fact that controlling environmental factors are highly complex in the two regions. There is the possibility that belt-like variations in hydrological conditions occur in the mountainous region due to varied relief and height-dependent variations in climatic conditions.

The present characteristics of the basic features of the hydrological cycle is by no means complete and only a broad interpretation is attempted. This is based on data collected along transects and on results of measurement made at three gauging points in the Tsagan-Turutuin-gol drainage basin in summer, 1975. The gauging points were installed:

a) at Olon-nuur at the outlet of a partial watershed (33 km^2) draining the central Khangai ridge,

b) at Mandal lying in the Tsagan-Turutuin-gol drainage basin I (1362 km^2) where the river emerges from the mountains,

c) at "Basin" point situated at the lower end of the Tsagan-Turutuin--gol drainage basin II (2180 km^2).

Between the Mandal and "Basin" gauging points the river crosses the wide Bayan-Nuurin-khotgor Basin, and no streams do contribute to its flow there.

All the Tsagan-Turutuin-gol catchment is underlain by discontinuous permafrost. The reported thickness of permafrost is 7—24 m, the active layer may be up to 3 m thick (Lonzhid 1966; Gravis 1974). In the mountainous part of the catchment, at 2100—2500 m asl., larch forest, 62 km^2 in area, occupies the north-facing slopes. Steppes and meadows are dominant, and tundra prevails on the mountain ridges. Cryoplanation terraces occurring in the summit areas, glacial cirques, lateral and frontal moraines and the flat bottom of the Bayan-Nuurin-khotgor Basin are of major importance in the hydrological cycle. Slope aspect is important locally for the storage of moisture. North-facing slopes are wetter than south-facing slopes. Differences vanish in the summit areas.

RUNOFF

Records covering the period 1 July—15 August, 1975, show that runoff values were above the average after a winter period of heavier than normal snowfall. Runoff values for 1975 are also high when compared to those for summer, 1974. At Mandal, the minimum discharge in the summer of 1974 was 0.67 m^3/s (Froehlich *et al.* 1975), while the corresponding value for 1975 was 5.0 m^3/s.

Table 3. Runoff values for the Tsagan-Turutuin-gol drainage basin
1 July—15 August, 1975

Drainage basin	Olon-nuur	Tsagan-Turutuin--gol I	Tsagan-Turutuin--gol II
Stream gauging point	Olon-nuur	Mandal	"Basin"
Elevation of gauging point [m asl.]	2650	2050	1950
Basin area [km^2]	33	1362	2180
Forest cover [%]	0	4.5	2.8
Runoff [million m^3]	2.010	80.42	67.06
Runoff [mm]	61.1	59.0	30.8
Q mean [m^3/s]	0.51	20.23	16.87
q mean [$l/s \cdot km^2$]	15.4	14.9	7.7
Q max [m^3/s]	2.35*	126.0	51.1
q max [$l/s \cdot km^2$]	115.0*	95.0	23.4
Q min [m^3/s]	0.15	5.0	5.25
q min [$l/s \cdot km^2$]	4.6	3.7	2.4

* Measurements taken above the lakes.

The characteristics of stream discharges for the Tsagan-Turutuin-gol catchment basin are given in table 3. Runoff values reported in cubic metres per period of record were 2.01 million m^3 in the Olon-nuur catchment, i.e. 2.5% of the runoff value at Mandal and 3.0% at "Basin". At the latter point the runoff value was by 13 million m^3 or 17% lower than that at Mandal. Channel flow here decreased downstream due to evaporation from the inundated plains, stream water supply to small lakes and water storage in the alluvial fill of the Bayan-Nuuring-khotgor Basin. Both runoff and mean diurnal discharges diminished downstream along the profile Mandal—"Basin", and at Mandal maximum discharges were 2.5 times lower than at "Basin" (tab. 3). Comparison revealed that minimum, mean and maximum specific runoff values for the Olon-nuur catchment were higher than those for the Tsagan-Turutuin-gol catchment at Mandal. The above variations are also demonstrated by estimates of the water storage capacity in the particular catchments (by using Maillet's formula) for

a) a period following three snowmelt floods (first half of July, 1975), and

b) a period following floods caused by excessive rainfall (first half of August, 1975).

The two types of floods that separated the periods of reduced channel flow were characterized by similar total runoff and maximum discharge values. The amounts of water which were gravitationally contributed to the streams during the first period were as follows: 3.6 mm (0.12 million m^3) at Olon-nuur, 1.77 mm (2.3 million m^3) at Mandal and 1.8 mm (3.9 million m^3) at "Basin". It appears that the amount of water per unit area which can be stored in the Olon-nuur catchment lying at higher elevations was twice as large as in the lower lying areas. Large amounts of water coming from the mountainous part of the Tsagan-Turutuin-gol catchment were also stored in the Bayan-Nuurin-khotgor Basin (i.e. below Mandal). During the second period of reduced flow the amount of water which here was stored was unimportant, while the reverse occurred in the Olon-nuur drainage basin.

It appears, that the Tsagan-Turutuin-gol catchment includes two regions of contrasting runoff volumes under the strong influence of environmental factors. The lower part of the catchment includes the sub-Khangai basin (below Mandal) and corresponds to Kuznetsov's region of decreasing runoff. It seems probable that during periods of very high flows this part of the catchment can be viewed as transitory with equal values of water input and losses. Data obtained from Olon-nuur and Mandal show that the southern part of the Khangai corresponds to Kuznetsov's region from which runoff is produced preferentially. The Tsagan-Turutuin-gol catchment I (above Mandal) is complex and comprises partial watersheds, a few tens of square kilometres in area, which provide ephemeral run-

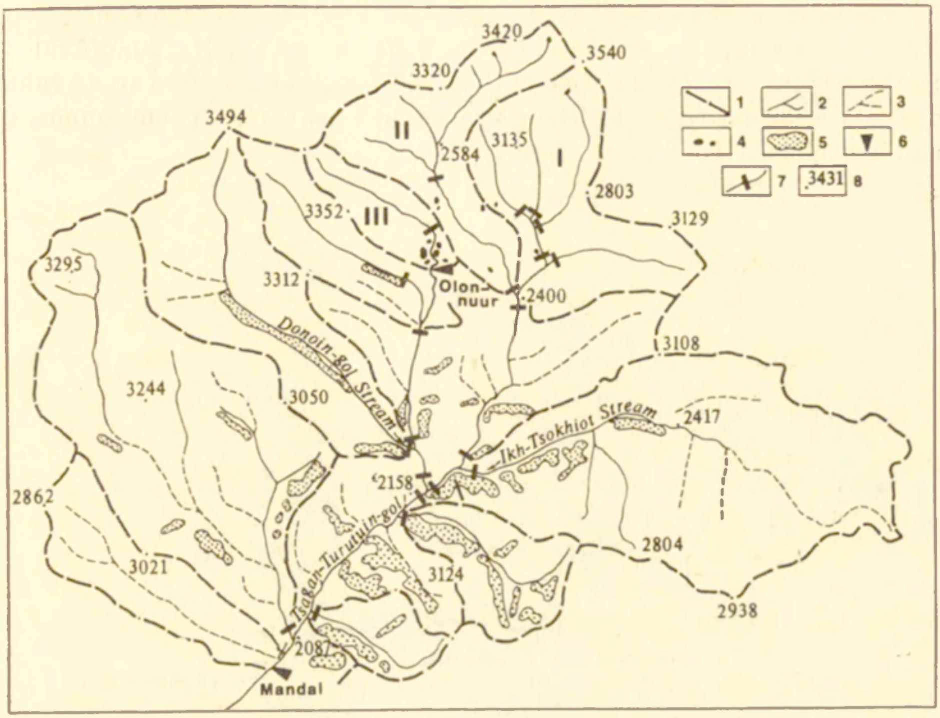

Fig. 11. The Tsagan-Turutuin-gol drainage basin (above Mandal)

1 — water divide; 2 — perennial stream; 3 — stream flowing seasonally; 4 — lakes and ponds; 5 — forest; 6 — stream gauging point; 7 — point of repeated measurements; 8 — altitude

off. Spatial variations in the runoff values from the mountainous Tsagan-Turutuin-gol catchment may be a reflection of the height-dependent changes in both precipitation and thermal gradients. In this part of the Tsagan-Turutuin-gol catchment superimposed on the climatically controlled variations are the influences of relief which are themselves reflected in the specific runoff values. These may vary with catchments of similar sizes, but (1) occurring either in the southern or northern part of the Tsagan-Turutuin-gol drainage basin and (2) differing by basin altitude. The control of both basin location and mean basin altitude has been confirmed by the results of discharge measurements. Specific runoff values ranged from 0—7.5 $l/s \cdot km^2$ on the southern mountain margin in partial watersheds covering 60—200 km^2 to 15—18 $l/s \cdot km^2$ in the northernmost, high-mountainous part of the Tsagan-Turutuin--gol drainage basin. Smaller partial watersheds (30—40 km^2) here provided perennial runoff, whereas to the south identical catchments produce runoff occasionally. It was found that in the small catchments which gained water from the central Khangai ridge the specific runoff values were related to basin size. For instance, in the catchments No. III, IV and V (fig. 11) covering 90—110 km^2 specific runoff values on 12 August, 1975, were 14.0—15.8 l/s · km². Discharge measurements made in their partial areas revealed by 30% higher specific runoff values. In the partial catchments greater than 150 km², the specific runoff values are less clearly related to basin size because of different environmental and hydrological conditions.

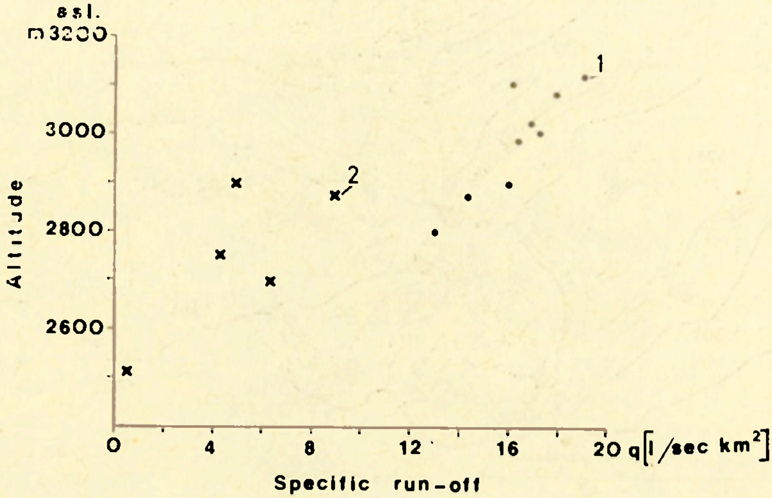

Fig. 12. Specific runoff value in relation to average basin altitude
1 — partial watershed receiving water from the central Khangai ridge; 2 — the other watersheds

In summer, by 50% higher specific runoff values were found in the partial watersheds that drain the main Khangai ridge having a glacial relief (fig. 12). For instance, specific runoff values on 6 August, 1975, ranged from 15.4 l/s · km² in the glacially modified Donoin-gol watershed (300 km²) to 7.4 l/s · km² in the Ikh-Tsokhiot watershed (275 km²) showing no traces of glaciation. The lower specific runoff values here reflect the interaction of lower altitude (fig. 11), location in the intermediate part of the mountains and a lack of water bearing glacial deposits.

The general pattern is one where

1) the highest specific runoff values correspond to the central Khangai ridge including the uppermost part of the Tsagan-Turutuin-gol drainage basin,

2) by 30—50% lower specific runoff values mark the intermediate part of this catchment with a perennial stream flow,

3) the lowest specific runoff values occur in the lower part of the

Tsagan-Turutuin-gol drainage basin (above Mandal) with both perennial and ephemeral streamflows.

In summer, the water balance is positive in the Tsagan-Turutuin-gol catchment extending above Mandal and negative below Mandal.

SURFACIAL HYDROLOGICAL PHENOMENA

The glacially modified relief of the mountains, permafrost occurrence, high local relief (up to 1500 m) and steep gradients combine to produce the areal variability of the surfacial hydrological phenomena in the Tsagan-Turutuin-gol drainage basin.

Within the Bayan-Nuurin-khotgor Basin which contains the lower part of the Tsagan-Turutuin-gol catchment examined (below Mandal) with a negative water balance there occur small lakes and swampy areas associated with streams. Ephemeral lakes occupying small thermokarst depressions are formed by thawing of frozen ground (Babiński and Grześ 1975). The hill- and mountain-sides of marked relief (to 800 m) which surround the inland basin are dry. Footslopes may be incised by small valleys owing a great deal to storm rainfall. In the Bayan-Nuurin-khotgor Basin no stream water is supplied to the main channel, and no ephemeral springs do occur there. Hydrological phenomena vary markedly within the mountainous part of the Tsagan-Turutuin-gol drainage basin I (above Mandal). In the floors of the deep and narrow valleys icings may persist until summer. In the main valley icings are found to 2900 m asl. Thawing icings supply water to seasonally wet grounds occupying valley floors. In the lower part of the catchment, perennialy wet grounds are due to ground water outflow, while in its upper part such grounds are associated with water flowing downslope from the flat mountain tops. The glacially modified valley floors in which grass grows luxuriantly have moisture in excess, and glacial cirques may be occupied by lakes. Wet grounds also occur in the mountain ridges at elevations exceeding 2600 m asl. On the high central Khangai ridge hydrological phenomena are arranged belt-like.

Spatial variations are discussed along a transect from the Donoin--Dzun-Nuruu ridge (3350 m asl.) to the Olon-nuur valley (2650 m asl.; fig. 13). The flat summit area which receives great amounts of precipitation is rather poorly drained because of its varied microrelief, and there is a tendency for water to collect in the form of ice. Consequently, depressions amid stony accumulations tend to be filled with water in summer, even during 2—4 mm rainfalls. Water is overflowing from one depression down to another. Because of this high storage property of the summit plateau downward movement of the water to deeper sites may continue for several days. Thus water is supplied to the mountain- and valley-sides. These also receive water from blockfields (fig. 13) with

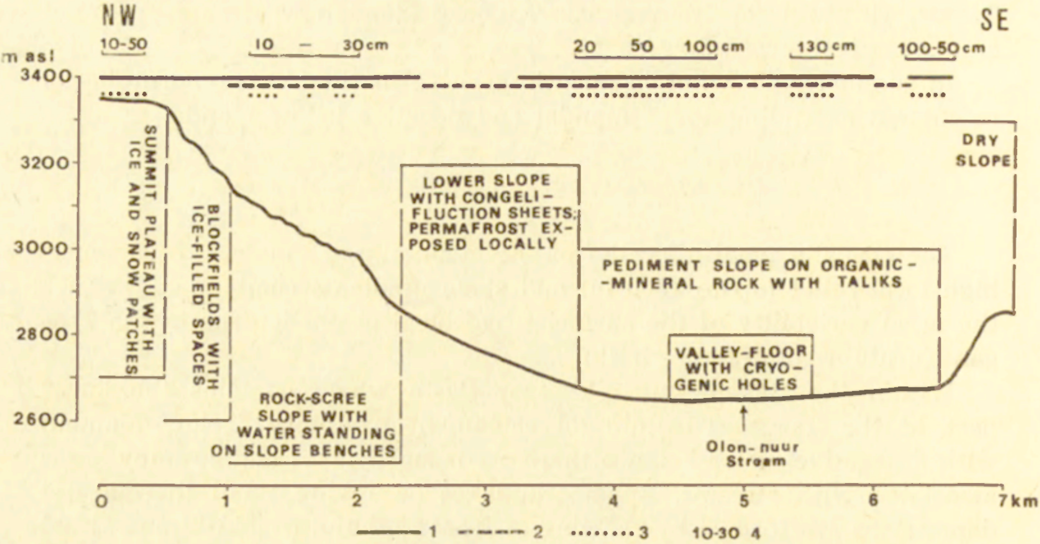

Fig. 13. Transect from the Donoin-Dzun-Nuruu ridge (3350 m asl.) to the Olon--nuur valley (2650 m asl.)

1 — surface runoff; *2* — through flow and interflow; *3* — cryogenic holes; *4* — depth to water level in the cryogenic holes (in cm)

spaces that contain winter ice melting throughout the summer. At lower elevations (below 3100 m asl.) water running downslope tends to saturate the uppermost thawed layer of the waste sheets. Thus small ponds may be produced on the slope benches. Intense creep which affects the water saturated slope deposits may inhibit perennial drainage formation. Traces of shallow stream channels are found below 2800 m asl.

On the smooth, sloping surfaces which extend below the steep valley--sides cryogenic holes [3] are widespread. These may occupy 30% of such surfaces. The depths of holes are shown in figure 13. It appears that the cryogenic holes are clearly important in the hydrological cycle. Through the cold season when soil temperatures fall below zero these holes are single, closed reservoirs. Through the summer meltwater trapped in the holes infiltrates slowly the peat and clayey sheets making its way to the lower holes and, finally, to streams and ponds which occupy the closed depressions. During the thaw season, i.e. at the beginning of July, the depths to water level within the holes varied from 0 to 50 cm. After rainfall which occurred in August the cryogenic holes were completely filled with water. Its discharge took place in a uniform manner. It appears that the storage capacity of areas comprising cryogenic holes is high in contrast to other areas.

[3] The word is Polish and means small vertical-sided pits probably formed by upfreezing stones in the active layer.

In the lowermost part of the transect discussed (fig. 13) there occur shallow, drained ponds which may store only 250 000 m³ of water in the Olon-nuur catchment. The south-facing slopes here are completely dry, and no traces of running water are present, while on the opposite slopes at 2800 m asl., the soil is water saturated. Such marked contrasts occur on the central Khangai ridge to elevations of 3000 m asl.

The distribution, discharges and temperatures of springs do not vary with height. Because of shallow permafrost only a few springs are found in the mountainous part of the Tsagan-Turutuin-gol drainage basin. Ephemeral springs predominate. Most of the springs emerge in the lower slopes of both alluvial fans and lateral moraines. In the mountainous part of this catchment hot springs (20—55°C), rich in sulphur compounds, emerge along the fault lines.

VERTICAL VARIATIONS IN WATER TEMPERATURES

The temperatures of stream water were recorded continuously at three gauging points. It appears that the monthly averages and the lowest values of the mean diurnal water temperatures in July, 1975, were always lower than the corresponding values of air temperatures along the whole Tsagan-Turutuin-gol river (tab. 4). E. Brzeźniak (1977) found that the

Table 4. Mean monthly air temperatures (according to E. Brzeźniak 1977) and corresponding water temperatures [°C] in the Tsagan-Turutuin-gol drainage basin

Drainage basin, July 1975			
Stream gauging point	Olon-nuur	Mandal	"Basin"
Elevation [m asl.]	2650	2050	1950
Mean monthly air temperature	8.2	12.6	13.4
Mean monthly water temperature	10.0	13.0	14.9
Highest mean diurnal air temperature	13.5	18.7	19.7
Highest monthly water temperature	14.8	15.4	17.4
Lowest mean diurnal air temperature	2.8	6.5	6.6
Lowest mean diurnal water temperature	ca. 3.5	7.6	9.6

air temperatures were closely correlated with the elevation above the level of the sea. The water temperature/altitude relationship is more complex. The height difference between Mandal and Olon-nuur is 600 m, and the variations in mean monthly water temperatures are 3.0°C. The height difference between Mandal and "Basin" is 100 m, and the variation in mean monthly water temperatures is 1.9°C. The river lengths between Olon-nuur and Mandal and between Mandal and "Basin" are essentially similar. In the mountains the river receives cold water, but in the piedmont basin it is warming up rapidly. Water temperature tends to exert a strong influence on the thermo-erosion of the river banks. Particularly in the Bayan-Nuurin-khotgor Basin, where conditions are most suitable, the above process is very intense (Klimek 1975).

In general, stream water temperatures at Olon-nuur were about 2—3°C higher than at adjacent sites. The cause was warmer water received from the cryogenic holes. The temperature of the topmost water layer here resembles more ground surface than air temperatures. In the cryogenic holes the warming up water produces "warm water pockets" within the active layer and causes a rapid thawing of permafrost.

Slope water temperatures recorded along the transect from the Donoin-Dzun-Nuruu ridge to the Olon-nuur valley were highest between 2900 m and 3100 m asl., and decreased to zero at lower elevations where the slopes exhibit congelifluction lobes.

CONCLUSION

In the Tsagan-Turutuin-gol catchment the water balance is positive in the mountains and negative in the mountain foreland. The line of demarcation corresponds to the morphological boundary of the mountains. Summer floods control the total annual runoff. In the mountain foreland covering some 800 km^2, runoff diminishes then by 20%. It is probable that water losses will be far greater downstream.

In the mountainous part of the Tsagan-Turutuin-gol drainage basin there height-controlled zones have been delimited.

In the highest, internal zone both temperatures and precipitation amounts favour perennial runoff production. This zone comprises partial watersheds with mean altitudes exceeding 2800 m asl. The watersheds occupy about 32% of total basin area extending above Mandal and only 17% of basin area extending above the "Basin" gauge. During the period of record these partial watersheds contributed 60% of the total runoff.

The intermediate zone includes partial watersheds of mean altitudes of 2500—2800 m asl. These watersheds may produce ephemeral runoff. They occupy 42% of total basin area (above Mandal) and contribute 30% of total runoff.

The lowest, external zone comprises partial watersheds of mean altitudes to 2500 m asl. These watersheds occupy 23% of basin area, but contribute only 10% of total runoff. Such watersheds occur in the marginal part of the mountains with highly varied hydrological conditions. South-facing slopes are extremely dry, whereas north-facing slopes are moister and forest-covered. Partial watersheds, up to 100 km^2 in area, are drained seasonally. In the drier years the hydrological cycle in this part of the Tsagan-Turutuin-gol catchment can be similar to that in the mountain foreland having a clearly negative water balance.

In the southern Khangai, the varied hydrological conditions show a belt-like arrangement which reflects the altitudinal zones of both thermal and precipitation conditions (Brzeźniak 1977) and of permafrost occurrence (Gravis 1974). The altitudinal zonation of hydrological conditions is reflected in the variability of both runoff values and surfacial hydrological phenomena.

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VI. PRESENT-DAY SLOPE MODELLING IN THE SOUTHERN KHANGAI MOUNTAINS

On the southern slope of the Khangai the morphodynamic belts correspond to the bioclimatic levels. Within these belts slope asymmetry is important in controlling the assemblages of morphogenetic processes. Both speed and course of the processes depend on climatic conditions and on the varied structure of bedrock because granitic and metamorphic rocks as well as basalt caps and flows are present in each of the belts.

From the height-dependent march of thermal phenomena (Brzeźniak 1977), the distribution of present-day permafrost (Gravis 1974) and the altitudinal zones of vegetation (Karamysheva and Banzragch 1977) which control the belt-like arrangement of both physical and chemical processes (Richter *et al.* 1963; Haase *et al.* 1964; Pękala and Ziętara 1977) on the southern slope of the Khangai the following distinctive morphoclimatic belts can be derived:

1) the belt of stony high-mountain tundra (above 3000 m asl.) and meadows (2700—3000 m asl.), with continuous permafrost;

2) the forest-steppe belt (2100—2700 m asl.), with discontinuous permafrost on north-facing slopes and strong process contrasts depending on slope aspect;

3) the steppe belt (below 2100 m asl.), with permafrost islands in the valley floors.

The position of boundaries among the different morphoclimatic belts is controlled by slope aspect.

PRESENT-DAY MORPHOGENETIC PROCESSES IN THE SUMMIT AREA

Summit plateaus attain a width of several kilometres and extend down from 3500 m to about 2400 m asl. in a series of steps. Both rounded ridges and narrow crests occur at lower levels. Structural control is expressed in the accordance of ridges with the jointing of bedrock.

The granitic sub-watershed plateau of the Donoin-Dzun-Nuruu is about 2000 m wide and to 4000 m long. It is slightly convex and smooth (3°) and has an extensive cryoplanation plain with patterned ground forms (sorted circles). These indicate that at present frost sorting is taking place in the active layer. Its thickness in July 1977 was 1 m. The mesh of sorted polygons is irregular, and stones predominate. The central areas have a concentration of clay with abundant granular waste and stones (2—20 cm in diameter) and bear a primitive brown soil. This is covered with moss and lichens (Kowalkowski 1977). The sorted polygons vary in size. The mesh size of smaller forms is 4 m across. The bordering tabular stones tend to dip at various angles being largely in the vertical plane. In general, stones are forming the bordering depressions.

Frost action processes contribute to size sorting of both blocks and fine particles in the waste because of repeated freezing and thawing. This fractional sorting involves upward movement of fines which may include clay-size particles. Thus circular, clayey with fine debris islands are formed amid blocks of varying size. Such forms have been described by Jahn (1975) as embryonic patterned ground. Even if embryonic forms occur in the Khangai, different stages in their development may be recognized.

The summit cryoplanation plain on granite passes into slopes with a series of cryoplanation terraces. The summit cryoplanation plains on basalt are almost flat. They also have sorted patterned ground forms. Their mesh is more regular than that on granite. The size range of stones, i.e. of columnar joint blocks is similar (0.5—1.5 m). The size range of fines forming the central areas also is similar. Nearly 30% of the mass are rock fragments measuring 0.5—1.0 cm across.

These sorted polygons owe a great deal to cracking caused by the freezing strain of basalt. On closer analysis the cracks are seen to be located along joints. Further frost sorting in the contraction cracks leads to the formation of sorted polygons because cracks favour concentration of sorted blocks (Jahn 1975). On the cryoplanation terrace edges the sorted polygons are less well developed, and contraction cracks are present. These are the forerunners of polygons. It seems that sorted polygon formation here is retarded by drier conditions (lower water level, free drainage of the active layer by valleys).

On the basalt-capped summit plateaus, cryoplanation flats are separated by edges from the cryoplanation terraces occurring at lower levels. On north-facing slopes the scarps are marked by frost-riven cliffs. At their bases snow-banks tend to persist until July. It is likely that present-day nivation processes cause the extension of the flats.

It appears that in the interior of the Khangai processes of ground swelling and frost cracking are most important in fashioning the summit plateaus. Intense physical weathering results in rock desintegration. In summer, chemical weathering attacks granite causing the decomposition of feldspar. The fines are carried away by eolian processes. Thus blow outs are produced on the clayey-blocks islands.

In the summit areas the processes discussed above are responsible for the formation of extensive cryoplanation plains being stepwise arranged. These developed within the older planation surfaces (Selivanov 1972; Sukhodrovsky 1975), and grade into the mountain- and valley--sides having convex stepped profiles. Cryoplanation flats on granite and metamorphic rocks are discordant features, whilst on the basalt caps they approximate to primary structure (accordance). On dissection, the horizontal fissures and columnar jointing of basalt determined the formation of scarps and edges which separate the summit cryoplanation flats from the slopes.

ALTITUDINAL ZONALITY OF SLOPE FORMS AND SLOPE SHEETS

In the interior of the Khangai, within the stony mountain tundra belt the development of slopes under periglacial conditions is related to primary slope form and mechanical composition of the waste covers. Within this belt the following slope varieties can be distinguished:

1. Steep walls enclosing glacial cirques and troughs bear thin, blocky--scree sheets in their upper portions. The overburden tends to thicken downslope. Below the walls materials derived from upslope are accumulated.

2. Mature slopes on granite have concavo-convex profiles. Slopes on basalt are straight in plan and pass with a concave sector into the footslope. Such slopes were affected by very strong periglacial influences during the Pleistocene valley glaciations. In the granitic area slopes are mantled with scree and clay. In the basaltic area slope sheets consist of blocks and scree.

Within the forest-steppe belt, slopes are classified into the following types:

1. Mature, asymmetrical slopes being concavo-convex in plan occur in the east-west trending valleys. On south-facing slopes scree and blocky sheets were encountered, whereas north-facing slopes bear clayey-block and blocky sheets. These occur mostly on upper slopes.

2. Steep slopes undermined by running water are either with or without a thin block and scree cover increasing in thickness on the lower slope.

The steppe belt comprises:

1. Straight slopes of tectonic origin bearing blocky-scree sheets. Such slopes are locally rejuvenated because of local base level changes.

2. Slopes resulting from fluvial erosion have sheets which consist

either of clay and scree or of clay (Kowalkowski *et al.* 1977). Such sheets occur commonly on metamorphic schist. Structure-controlled single tors. scarps and edges are found on the cryopediments.

PRESENT-DAY SLOPE MODELLING IN THE MOUNTAIN TUNDRA BELT

The steep walls of both glacial cirques and throughs are fashioned by physical weathering processes and mass wasting. Consequently, the uppermost parts of the rock walls become smoother (46—58°), and the lower parts are built up by talus. The stony waste is removed by dirty avalanches in the spring and by rainwash in the summer. The frozen ground favours sliding of the rocky debris. Bedrock striations and grooves occurring on the mid-slopes indicate intense processes of corrasion. Both separation and removal of slabs of rock is controlled by jointing of the bedrock.

The lower scree slopes are modified by rapid sliding of dirty avalanches which erode tracks a few metres deep. Ridges developed beside the tracks are characteristic features of avalanche deposition.

By stripping off scree and blocks mass wasting reduces the slope gradient to 34—48°C, and a concave sector occurs only at the foot of slope. Below the north-facing slopes debris accumulations are found. The debris is thought to have moved down over the snow patches which tend to survive in sheltered sites.

In the interior of the Khangai, the fashioning of the majority of slopes on granite and basalt is atributed to dominant processes of physical weathering, various kinds of gravity mass movement (largely congelifluction), deflation, washing out of fines and slope wash (Ziętara 1976a) occurring under past (Pleistocene) and present periglacial conditions. At present the slopes that are underlain by permafrost have reached the mature stage in their development.

The concavo-convex, stepped slopes of the Donoin-Dzun-Nuruu are a fine example of mature periglacial slopes on granite (fig. 14). On the upper slope cryoplanation terraces occur. The mid-slope has stone-banked congelifluction lobes, and on the lower slope both congelifluction lobes and terraces are found. The higher jointcontrolled cryoplanation terraces sloping at 14—18° outward attain widths of 100—400 m. The lower cryoplanation terraces are wider, while gradients are lower (to 8°). On these flats frost sorting gives rise to stony circles which are frequently elongated parallel to slope tread.

The terrace scarp edges of 25—30 m have "imbricated" blocks of varying size. These are attacked by physical weathering processes (mostly exfoliation) producing various microforms. Granular waste is also formed. Such processes are most important in the spring and summer. The waste is swept away by rain.

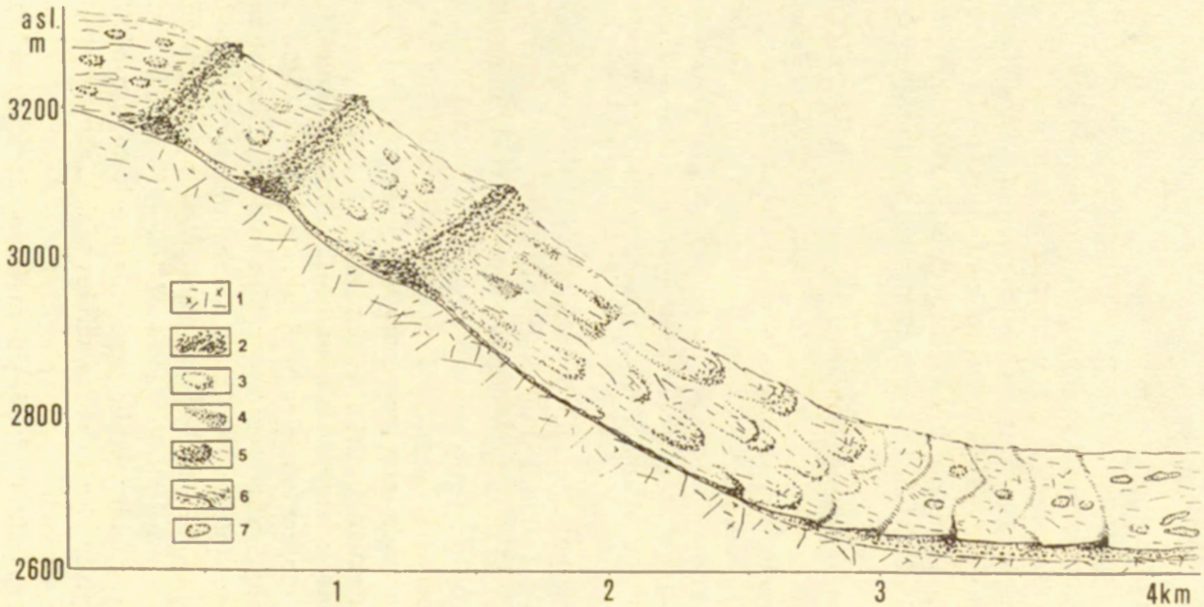
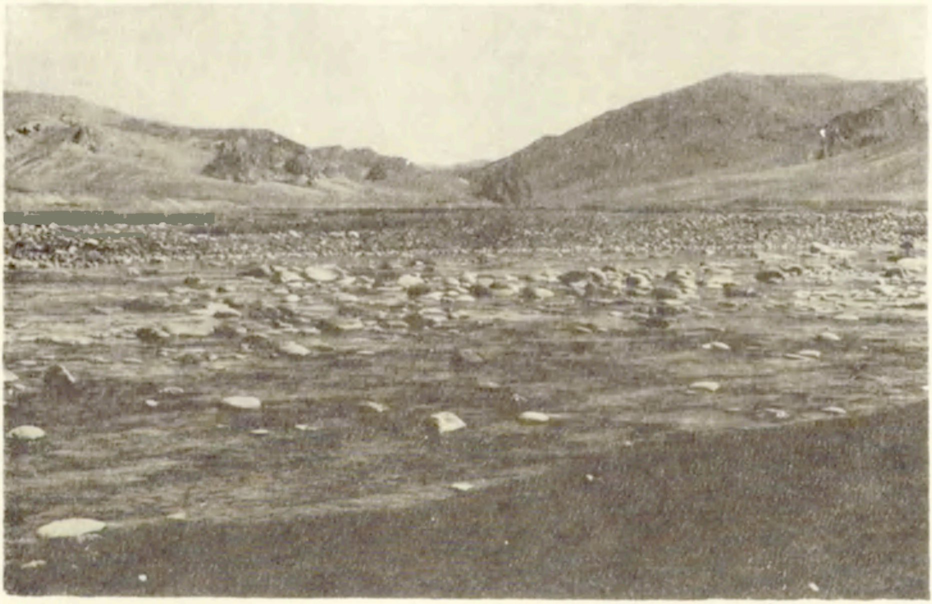


Fig. 14. The concavo-convex slope of the Donoin-Dzun-Nuruu fashioned under periglacial conditions the Khangai (by T. Ziętara)

1 — granitic bedrock; *2* — cryoplanation terrace edges on angular granite blocks; *3* — sorted polygones; *4* — stony stripes; *5* — congelifluction lobes; *6* — congelifluction terraces; *7* — cryogenic "windows"

Phot. by K. Pękala

Phot. 1. The extensive summit plateau developed on the central Khangai ridge (3300—3400 m asl.) has a scare high-mountain tundra vegetation cover and numerous patterned ground forms. Higher cryoplanation terraces with marked edges and glacial cirque walls are seen in the far distance



Phot. by W. Froehlich

Phot. 2. In the marginal zone of the Khangai the wide floor of the Tsagan-Turutuin-gol valley is strewn with coarse alluvia. In the distance the steep valley- and mountain-side are occupied by the mountain steppe

Phot. by K. Klimek

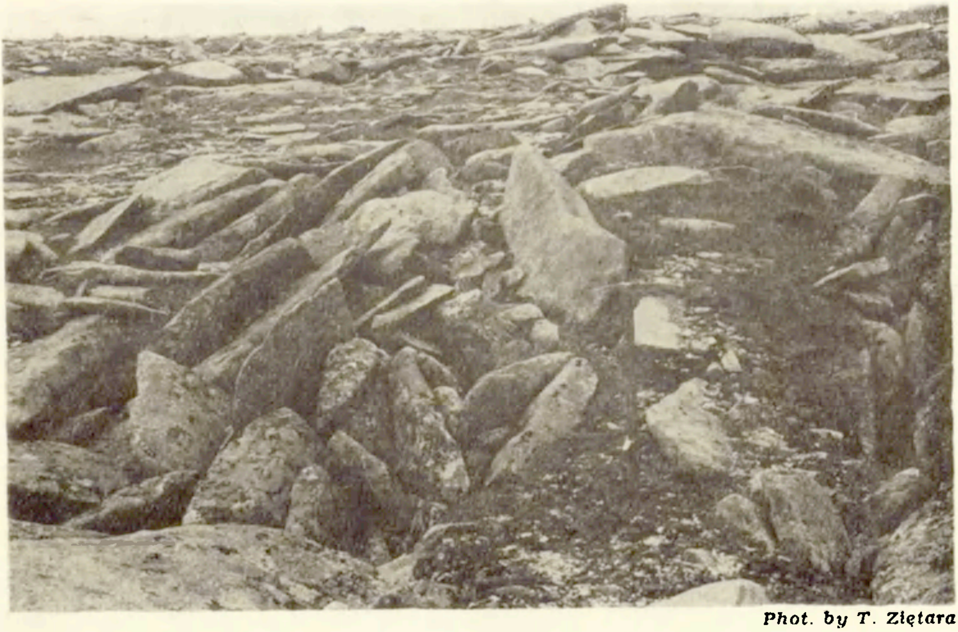
Phot. 3. Upstream view of the Tsagan-Turutuin-gol valley, where it emerges from the mountains. Observe the well preserved edge of the Pleistocene terrace bearing slope sheets (right) in the sub-Khangai rift-valley

Phot. by K. Klimek

Phot. 4. A roche mountonne in the floor of a formerly glaciated valley in the Khangai. The blocky lateral moraine occurring in the background is dissected by ravines, now dry

Phot. by T. Ziętara

Phot. 5. A cryoplanation terrace edge with imbricated granitic block



Phot. by T. Ziętara

Phot. 6. Sorted polygon on a cryoplanation terrace occurring on the central ridge (internal zone) of the Khangai

Phot. by K. Pękala

Phot. 7. An interfluve in the densely dissected marginal zone of the Khangai

Phot. by K. Pękala

Phot. 8. Frost-riven cliffs are rising above the fossil cryoplanation terraces in the marginal zone of the Khangai at 2700 m asl.

Hydrological conditions are of great importance in the fashioning of the mid- and lower slope sectors by congelifluction lobes and terraces. In the summer of 1975 permafrost occurred at depths of 0.5—1.2 m. Seepage of meltwater from the terrace edge bases facilitated the formation of stony stripes and lobes which spread fan-like, and of lobes with marked frontal stone banks.

The lower, concave-stepped slope sector is being fashioned by congelifluction producing crescent-shaped terraces and garlands. The clayey, vegetation-covered terrace surfaces have cryogenic "windows" [4] (Zietara 1971) and debris islands. These holes are the sites for collection of moisture and thus the sites of most intense frost action.

The development of slopes on basalt slightly different from that in granitic areas. In the upper slope sector cryoplanation benches and terra-

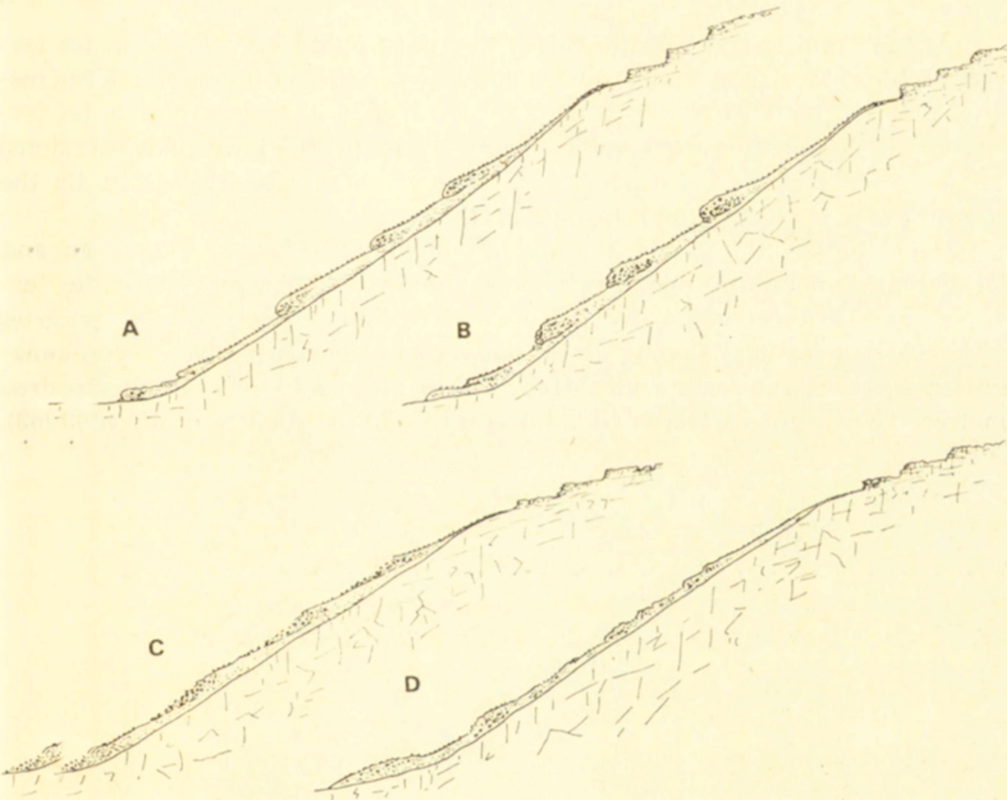


Fig. 15. Stages in the development of a mature slope on basalt, the Khangai (by T. Ziętara)

A — sod-covered congelifluction lobe; B — downward movement of the lobe is arrested by the growing stone bank; C — rupture of the vegetation cover releases a flow of debris which levels the slope; D — formation of stone-banked lobes

[4] See note 3, page 48.

ces are found. Terrace surfaces slope gently at 8—14°. The mature slope on basalt is stepped and straight in plan (fig. 15). The footslope is concave. Congelifluction terraces occur locally where the displaced masses contain about 50% of clay. On the basaltic slope congelifluction lobes tend to develop, and constricted congelifluction occurs there (fig. 15 *A*). The length of lobes overriding humus-vegetation covered surfaces may reach 50 m. Stones may move downslope faster than slow flowage of the fine materials forming frontal banks (fig. 15 *B*). The downward movement of the masses is arrested by the growing stone bank. The sod cover may be ruptured in places releasing a flow of fine debris of similar size. This debris tends to smooth the stepped slope profile (fig. 15 C). Slope wash also can level the slope steps. Rock particles are further comminuted by frost action and repeated mass movement produces small stone banked terraces (Jahn 1975). At later stages stone-banked lobes give rise to a stepped slope being straight in plan (fig. 15 *D*).

On the granitic and basaltic ridges there are found cryoplanation terraces which developed under severe climatic conditions (Boch and Krasnov 1951; Reger and Pewe 1976). The occurrence of cryoplanation terraces in both an active state and an inactive state (at the lower elevations) in this area strongly suggests that changes of climate took place in the immediate past (Kowalkowski *et al.* 1977).

Cryoplanation terraces are found on mountain-sides on ridges and on flattened mountain-tops which extend above 2400 m asl. These terraces are best developed above the upper tree line and comprise a series of bench-like features sloping at 12° outward to the scarp edge. Cryoplanation terraces range in width from a few metres to several hundred metres. The height of scarps (fig. 16) is up to 75 m (Richter *et al.* 1963).

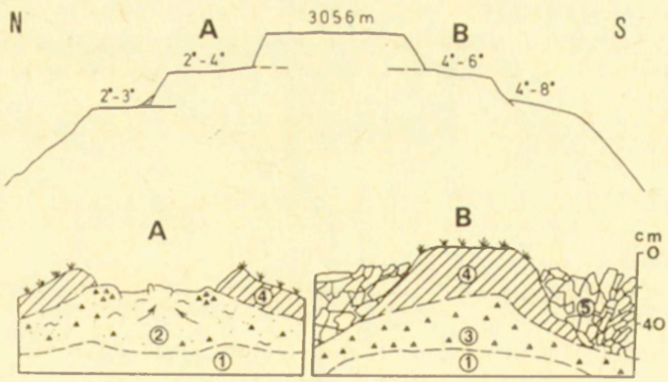


Fig. 16. Scheme of cryoplanation terraces developed on a residual hill, the Delgerin-Ikhe-Nuruu massif (by K. Pękala)

A — patterned ground on a north-facing slope; *B* — patterned ground on a south-facing slope; *1* — permaforst layer; *2* — clay with fine angular rock fragments, highly water saturated; *3* — fine, angular rock fragments; *4* — soil (humus horizon); *5* — stone circle

PRESENT-DAY PROCESSES OCCURRING IN THE FOREST-STEPPE BELT

The forest-steppe belt developed near the lower margin of the Khangai. Slope asymmetry is distinct in granitic and basaltic areas. Discontinuous permafrost underlies forest growing on north-facing slopes (Pękala and Ziętara 1977).

The fashioning of south-facing slopes is largely influenced by physical weathering. This is accompanied by sheet- and rill-wash, whilst creep of the active layer is significant on north-facing slopes. On the granitic, south and southwest-facing slopes selective weathering and removal of waste products is responsible for the formation of numerous structure--controlled benches, tors, and even of hard-rock ridges (fig. 17). On basaltic slopes, processes oriented laterally permit slope foot erosion at the level of flat-floored valleys. This levelness extends headwards (fig. 18) and the resultant slope is concavo-convex in plan.

The mid-sectors of north-facing slopes have smaller gradients (to 32°) than those of the south-facing slopes (38—42°). On the latter slopes numerous tors occur. Deluvia and colluvia accumulate in the alley floors. The valleys that are drained by streams flowing only seasonally (e.g. the

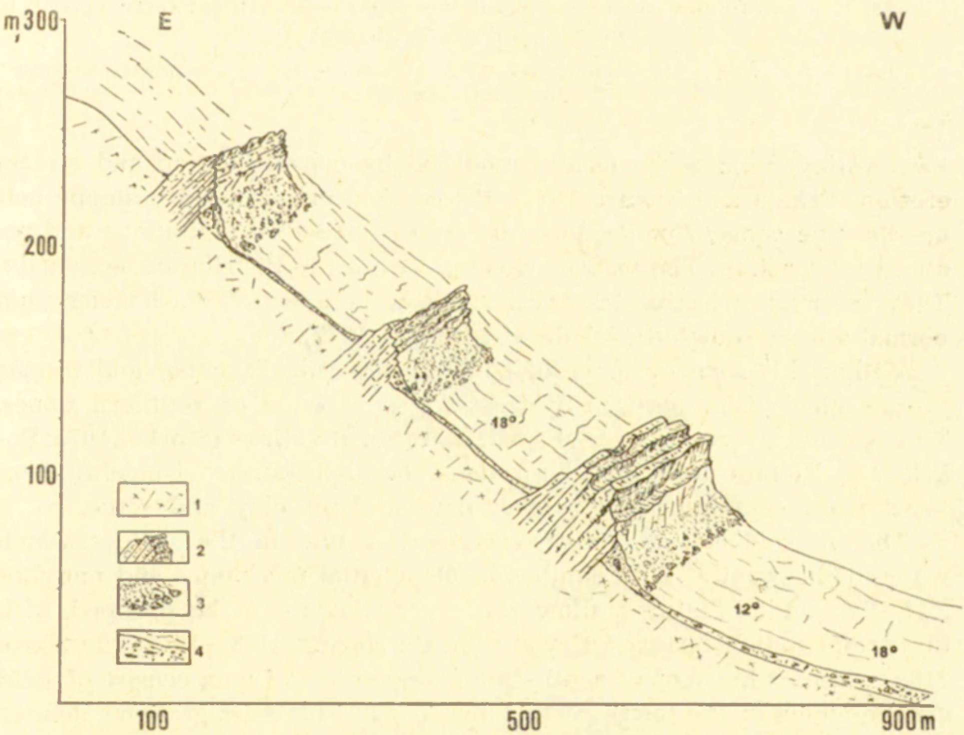


Fig. 17. Slopes on granite developed within the forest-steppe belt; the Khangai (by T. Ziętara)

1 — granite; 2 — hard-rock ridges on aplite veins; 3 — talus cones; 4 — waste sheets

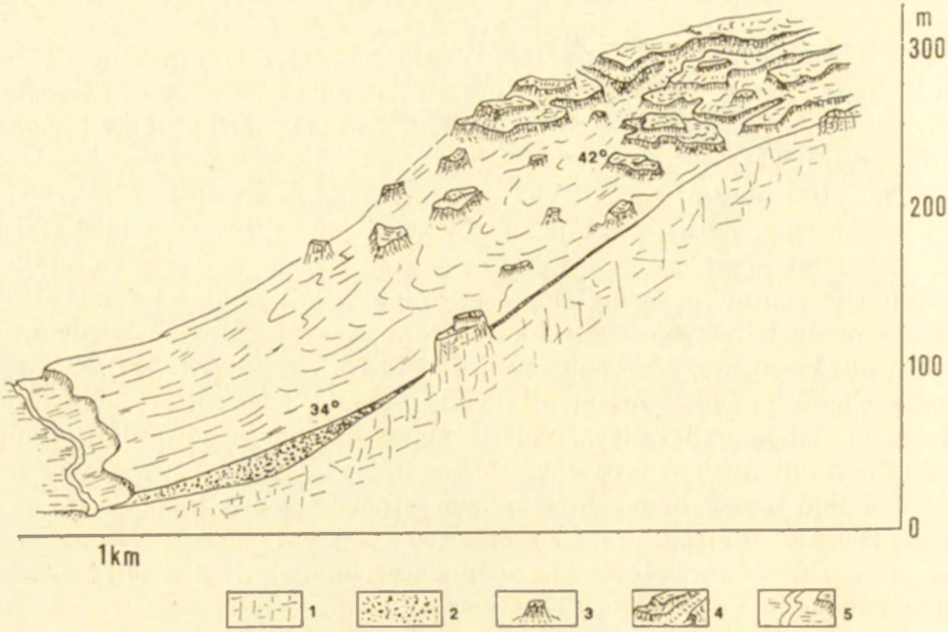


Fig. 18. Concavo-convex slopes on basalt developed within the forest-steppe belt, the Khangai (by T. Ziętara)

1 — basalt; 2 — waste and proluvial sheets; 3 — residuals (tors); 4 — avalanche tracks; 5 — stream channel

Sant valley) tend to be locally modified by congelifluction and stream erosion (Pękala and Ziętara 1977). Below 2300 m asl. the dry steppe belt has inactive congelifluction lobes because of moisture limitations and reduced frost action. The various slope processes are distributed seasonally. They are most effective in spring and wet summer after a heavier than normal winter snowfall (Pękala and Ziętara 1977).

Within this zone evidence of both former congelifluction and former intense block disintegration by frost action is found on southern slopes. This is deduced from saucer-shaped, corrasional valleys (Starkel 1975; Pękala and Ziętara 1977). In the valleys the stone-banked congelifluction lobes were overriding older lobes consisting of clay and scree.

The other evidence are cryopediments found in the valleys which were cut in granite. Both similar developmental mechanics and morphologically similar features allow the cryopediments to be grouped with the cryoplanation terraces. Cryopediments sloping at 5—12° occur above 2100 m asl. at the foot of south-facing slopes. These flats consist of solid granite which in the lower part of a cryopediment bears granular deluvia including buried chestnut soils (Kowalkowski *et al.* 1977; Starkel *et al.* 1975). For this reason the cryopediments are sub-fossil forms whose edges are marked by relict tors. There also occur stone streams, and colluvial-

-torrential fans have accumulated at the mouths of small, dry side-valleys. The cryopediments pass into the 8—15 m fluvial terrace plain which formed the base-level in Pleistocene times. It appears that at that time the climate was wetter and cooler than today to allow the development of cryopediments. The major part of their formation is attributed to frost, breaking up the bedrock and comminuting debris to smaller particles thus facilitating the downslope removal of the waste.

Within the forest-steppe belt, alteration of the relict moraines is taking place. Their flanks consisting of clay and blocks are refashioned by creep and ploughing blocks due to needle ice growth and water saturation of the ground. The downslope collapse of single blocks may also be the effect of piping, deflation and burrowing activities of animals. During the thaw season and heavy rainfalls the supersaturated granular waste with rock debris tends to flow downslope. On the flat moraine ridges the upheaving of blocks and sorting of moraine deposits produces polygons in moister sites, whilst in drier sites weathering causes the breakdown of upward projecting blocks.

PRESENT-DAY PROCESSES OCCURRING IN THE STEPPE BELT

Broad valleys and inland basins are widely distributed in this altitudinal zone. The bounding scarps are often of tectonic origin, with rockfall and seasonal gullying being accompanied by deposition of large torrent fans at the scarp foot. The actively uplifted tectonic scarps are affected by landslide movements related to intense earthquake activity (Ziętara 1976 b).

In general, the mature slopes are concave in plan. Their lower sectors have gradients of 10°, and the upper sectors are inclined at between 16—20° depending on geological structure and stage in development. On the upper slope numerous tors and structure-controlled benches are found. The footslope is mantled with sheets consisting of washed materials (Starkel 1975; Kowalkowski *et al.* 1977). These indicate that wash is important in the actual fashioning of the slopes. On the nearly bare surfaces this process is accelerated by grazing cattle and horses. Deflation has significance on shattered metamorphic rocks (Rotnicki and Lomborinchen 1977).

The broad basin and valley floors are underlain by discontinuous permafrost which influences the present-day morphogenetic processes. Frost action produces patterned ground forms which may occur in groups or single. Their formation is controlled by the composition of the basin and valley fills. Hydrological conditions are also important, for instance, the occurrence of hydrolaccolithes in oxbow-lake sediments (Rotnicki and Lomborinchen 1977).

RESULTS

The height-dependent differentiation in morphogenetic processes in the southern Khangai coincides with the occurrence of the altitudinal zones of climate, vegetation and soils.

Within the piedmont steppe belt, slope wash. seasonal stream erosion, piping and deflation are important in the fashioning of landforms at the present day. On moister terrace flats frost sorting, frost heaving and frost cracking may occur.

Within the forest-steppe belt, by far the most important process is shee- and rill-wash. Seasonally occurring stream erosion and deflation have greater significance in the summit area. Permafrost favours creep. Congelifluction, piping, corrasion and saltation by uprooted trees are of little importance.

Within the high-mountain meadow and tundra belt, periglacial process influenced by the presence of discontinuous permafrost are of major importance in the refashioning of the summit areas, of both mountain- and valley-sides and of valley floors. This assemblage of processes involves preparatory physical weathering, frost heaving, frost sorting and frost cracking occurring in association with congelifluction, wash, mass movement, nivation, wind action and corrasion. The effect are cryoplanation terraces and stepped slope profiles. In both summit areas and valley floors patterned ground forms tend to develop. At present such forms reached various stages in their development.

Slopes developed under periglacial conditions are irregular in plan and levelled locally by processes being oriented up- and downslope and laterally. Structural control is reflected in the slope form, slope angle varying with slope sector, and in the assemblage of different slope processes.

In the southern Khangai cryoplanation terraces are classified into the following types depending on the morphologically different features and the different dynamics of the present-day cryogenic processes: most active forms, active forms, less active, fossil forms and sub-fossil forms.

The most active cryoplanation terraces occur on the central Khangai ridge, within the mountain tundra belt which reaches above 3200 m asl. Cryogenic processes here are working areally and slope aspect is unimportant. Cryoplanation terraces are in the form of large summit flats with a scree-block mantle. This material is underlain by shallow permafrost, and a typical feature is the seasonally abundant moisture. On sloping ground debris removed from the cryoplanation terraces by congelifluction tends to accumulate above the intervening scarps thus producing stone-banked garlands. Piping removes the fines from the congelifluction sheets, and gravitation processes being aided by nivation transport the coarser material to the bottoms of glacier cirques and throughs. Debris also tends to built up congelifluction terraces on the slopes.

The active cryoplanation terraces are found on mountain-sides and on tops which lie south of the Continental Divide. On south-facing slopes these landforms occur above 2900 m asl., while those on north-facing slopes are recorded above 2800 m asl. On the latter slopes the cryoplanation terraces are inclined at 2—4°, and the rocky cliffs are largely vertical. The level surface is one of block- and clayey waste being modified by frost sorting processes. The central area of patternal ground — forms is clayey and mobile because of wetting.

The less active cryoplanation terraces which occur on north-facing mountain-sides and on tops rising above 2700 m asl. represent the transition forms from the active to the fossil terraces. At present only the scarps are attacked by weathering and the lower parts of terrace flats are modified by slow congelifluction and frost sorting.

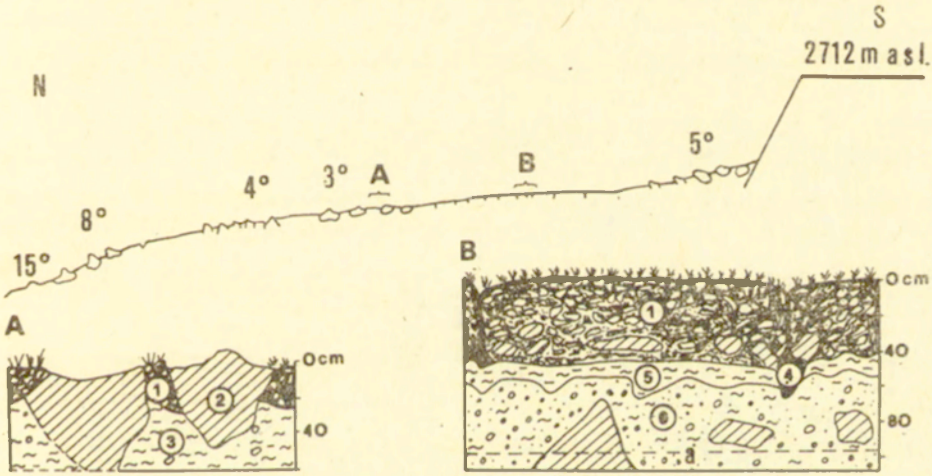


Fig. 19. Long-profile of a fossil cryoplanation terrace (according to Kowalkowski *et al.* 1977)

A — weathering causes the breakdown of upward projecting granite blocks; *B* — a patterned ground-form; *1* — humus-rich chernozem horizon with a pavement and floating sorted blocks; *2* — granite block attacked by weathering; *3* — clay with fine, angular rock fragments; *4* — humic infill of cracks; *5* — silty clay disturbed by frost action; *6* — clay with angular rock fragments and a granular waste; *a* — permafrost layer

The fossil cryoplanation terraces are found within the high-mountain meadow belt in the height range 2400—2700 m asl. These forms may be both inactive and slightly modified by frost action (Kowalkowski *et al.* 1977; fig. 19). Slope aspect played a distinct part in the development of different cryoplanation landform assemblages. On the northern mountain sides these forms comprise wide smooth surfaces and bounding steep scarps or tor-like rocky outcrops. On south-facing slopes the development of cryoplanation terraces was facilitated by locally deeper thawing of permafrost, by rapid removal of the coarse waste and by intense weathering of the exposed rock.

In the Khangai, slope aspect is also very important in the development of mountain- and valley-sides. Slope aspect controls the thickness of the active layer and the complex of present-day processes as well as their rhythm and intensity.

The southern Khangai together with its foreland is affected by young tectonic movements causing changes in the local base levels. These bring about changes in both slope profiles and further tendencies toward slope development.

ALOJZY KOWALKOWSKI

VII. ALTITUDINAL ZONATION OF SOILS IN THE SOUTHERN KHANGAI MOUNTAINS

The humidity of climate in the interior of Asia is taken as the major distinction between different structures and varieties of soil belts in the mountains (Dorzhgotov 1973; Stepanov 1975). However, the spatial distribution of precipitation amounts is locally modified in the various massifs.

On the southern slope of the Khangai soil formation depends not merely on the altitude and air moisture content, but also on slope aspect. In this paper an interpretation of soil belts in the southern Khangai is attempted, based on data collected by the present author on the Donoin-Dzun-Nuruu ridge and in the Sant valley (1974—1975), along with data contributed by other authors.

THE TIME ASPECT OF SOIL FORMATION

On the southern and western slopes of the Khangai several local varieties of soil belts have been recognized (Haase 1963; Kowalkowski and Lomborinchen 1975; Karamysheva and Banzragch 1976 a; Kowalkowski 1977). These differ from the semi-humid Khangai variant which was distinguished by Dorzhgotov (1973). The formation of the varieties of soil belts is directly controlled by the arid air masses which travel northward from the Gobi into the interior of the bordering mountains.

The soil cover does not correspond with the existing plant cover which reflects the actual climatic conditions. A similar conclusion was reached by Karamysheva and Banzragch (1976 a) in the Khan-Khudsiyn Ula massif. A suggested interpretation is that on the southern slopes of the Khangai soil formation was influenced by changes of both climate and plant communities with time. These produced mosaics and catenas with retardative, regressive and progressive features.

SOIL FORMING PROCESSES

A characteristic feature of the soils examined are both relict and present-day brown wastes of the periglacial environment incorporated

Table 5. Characteristics of some soil

Bv — brown waste; d — deluvial material; t — clay enrich-

Soil belt	Profile no.	Altitude [m]	Slope aspect	Depth of sample [cm]	Genetic horizon
Mountain primitive brown soils	037	3340	SE	0–5	BvC
				20–25	BvCg
Mountain brown soils — structure ground affected by congelifluction, underlain by permafrost	038	3340	SE	0–5	BvA
				10–15	Bvt
				25–35	BvCg
	010	2415	N	0–5	dBvtA
				20–25	dBvtA
				35–45	dBvA
				90–100	dBv
				135–140	dBv
Mountain grey-brown soils — structure ground affected by congelifluction, underlain by permafrost	039	3110	SE	0–5	dBvA
				25–30	dBvtA
				60–70	dBvtA
				50–60	dBvCg
Mountain chernozem soils — structure ground affected by congelifluction, underlain by permafrost	040	2950	SE	5–10	dBvA
	041	2810	SE	40–50	dBvtA
				0–5	dBvtA
				35–40	dBvtA
				50–55	dBv
				70–75	dBv
	009	2524	N	0–5	BvtAk
				25–30	BvtA
				40–45	Bvt
				70–75	Bv
Mountain dark chestnut soils, underlain by permafrost	043	2740	SE	5–10	dBvtAk
				30–40	dBvtAk
				50–60	dBv
	013	2476	N	0–5	dBvtAk
				10–15	dBvtAk
				40–45	dBv
				100–105	Bv
	014	2499	N	0–5	BvAk
				25–30	BvAk
	008	2522	S	0–5	BvtAk
				25–30	BvtAk
				55–60	BvA
				90–100	Bv
				110–120	Bv
Mountain light chestnut soils, poorly developed, denuded	006	2470	S	0–5	dBv/A/k
	004	2410	S	80–85	dBv/A/k
				0–5	dBvAk
					dBvk

features and soil properties

ment; A — humus accumulation; k — chestnut waste

Colour	C [%]	N [%]	C : N	pH_{KCL}	$CaCO_3$	Fractions [%] > 1.0 mm	Fractions [%] < 0.02 mm
2.5Y 5/4	0.55	0.07	7.8	4.1	0	18.2	33.5
2.5Y 5/4	0.43	0.06	7.2	4.0	0	17.1	24.9
10YR 2/2	5.81	0.85	6.8	4.6	0	17.7	7.4
2.5Y 4/3	0.55	0.07	7.8	4.4	0	15.6	22.8
2.5Y 5/4	0.35	0.05	7.0	4.4	0	20.3	20.7
10YR 4/3	4.37	0.35	12.4	5.5	0	5.8	35.8
10YR 6/4	1.67	0.12	14.5	4.7	0	26.6	30.8
10YR 6/4	1.08	0.08	13.8	4.5	0	57.4	10.6
10YR 5/8	0.41	0.03	12.0	4.4	0	51.7	9.7
7.5YR 5/8	0.28	—	—	5.0	0	56.7	4.8
10YR 3/2	6.18	0.47	24.1	5.1	0	0	22.0
10YR 3/3	2.59	0.29	8.9	4.5	0	6.9	27.0
2.5Y 4/6	1.90	0.07	27.1	4.4	0	10.1	29.6
2.5Y 4/3	0.59	0.07	8.4	5.1	0	25.6	17.1
10YR 2/2	8.90	0.89	10.0	4.4	0	7.8	24.9
10YR 2/3	7.61	0.76	10.0	4.0	0	6.6	34.6
7.5YR 2/2	9.57	1.06	9.0	4.7	0	1.3	38.5
7.5YR 2/1	7.95	0.86	9.2	4.3	0	13.6	31.9
2.5Y 4/4	2.04	0.20	10.2	4.0	0	24.2	18.9
2.5Y 4/4	1.44	0.14	10.3	3.9	0	13.1	17.2
10YR 3/2	5.20	0.50	10.3	5.6	0	11.6	41.9
10YR 3/2	3.90	0.46	8.4	5.5	0	13.7	37.3
2.5Y 5/4	0.89	0.10	8.6	5.4	0	25.1	25.8
2.5Y 5/4	0.72	0.09	8.1	5.4	0	33.5	21.5
2.5Y 3/2	4.68	0.67	7.0	4.9	0	1.4	33.5
7.5Y 3/3	1.89	0.26	7.3	4.6	0	4.7	37.2
2.5Y 4/4	1.25	0.12	10.4	4.6	0	32.5	15.5
10YR 4/3	6.62	0.57	11.7	5.4	0	6.2	21.6
10YR 5/4	2.40	0.19	12.6	5.0	0	30.4	22.9
10YR 5/3	0.88	0.08	10.0	4.8	0	71.9	9.8
2.5Y 5/3	0.29	0.03	8.8	4.7	0	22.4	10.1
10YR 3/4	4.70	0.40	11.8	5.6	0	9.8	32.5
10YR 3/4	1.12	0.13	8.7	5.3	0	65.8	13.7
10YR 3/3	4.42	0.35	12.6	6.0	0	11.6	26.5
10YR 3/4	2.96	0.39	7.6	6.4	0	13.7	30.2
10YR 3/6	1.13	0.24	4.8	5.4	0	25.1	23.2
10YR 3/8	0.41	0.04	9.5	5.6	0	33.5	14.0
10YR 3/8	0.36	0.06	6.2	5.6	0	33.1	16.1
10YR 4/3	1.65	0.13	12.3	5.2	0	31.1	13.1
	1.18	0.12	10.0	5.9	0	40.0	14.4
10YR 3/3	2.45	0.27	9.2	5.6	0	22.1	11.7
10YR 4/6	0.99	0.12	8.0	5.7	0	36.2	12.1
	0.45	0.06	8.2	5.9	0	40.7	11.3

in the soil profiles independently of altitude, slope aspect and precipitation—evaporation ratio (tabl. 5, prof. no. 037, 038, 039, 041, 008, 043, 013). The soils are acid (pH 6.0—4.0) down the whole profile and lack carbonates and salts (tab. 5). These features point to a former wetter and cold climate (Kowalkowski 1975; Nogina 1978).

The dominant role played by temperatures in both present and past soil formation is indicated by the occurrence of mechanical foliation planes on the surfaces of quartz grains in structure (patterned) ground on the Donoin-Dzun-Nuruu ridge, and of thermal disintegration in the chestnut soil which occupies steppe lands in the Sant valley (Kowalkowski and Mycielska-Dowgiałło 1980. These are also found in the whole proluvial series of profile no. 023 including the buried chernozem at base (Kowalkowski *et al.* 1977). In the structure ground associated with the active layer (prof. no. 037 and 038) processes of chemical weathering of the quartz grains and of siliceous crust formation were observed, while steppe soils and the buried chernozem exhibit only traces of a relict chemical weathering. Such traces are absent from the series of proluvial deposits whose dominant features are those of thermal jointing and frost exfoliation of both grain surfaces and relict siliceous crusts as well as the precipitation of silica from the freezing and evaporating soil solutions that are migrating upward in the dry and cool continental climate.

Within the present forest-steppe belt, chemical processes take place largely under the influence of capillary moisture rising in soils that are underlain by permafrost, i.e. on north- and west-facing slopes. On the contrary, on the south- and east-facing slopes processes of mechanical degradation and aggradation are most important. These affect both slopes and soils and are accompanied by intense temperature-dependent processes of physical weathering, soil creep, slope wash, deflation as well as fluvial erosion and deposition (Pękala and Ziętara 1977).

On the Donoin-Dzun-Nuruu ridge, water is important in the migration of both mineral components and organic matter, and in the transfer of soil materials on the slopes. These materials have characteristic autonomic features within the particular soil belts.

THE STRUCTURE OF SOIL BELTS IN THE SOUTHERN KHANGAI

The height-dependent hydrothermal conditions are the prime cause of the development of soil belts. Clear boundaries exist among belts that developed within the cryo-humid [5] zone extending above altitudes of 2500—2700 m (fig. 20). Within the lower, cryo-arid zone such boundaries are less distinct. Soil belts were distinguished on the base of the major soil types which appear to be controlled by altitude, slope aspect, and

[5] The words "cryo-humid" and "cryo-arid" are Russian and mean either wet frozen or dry frozen ground.

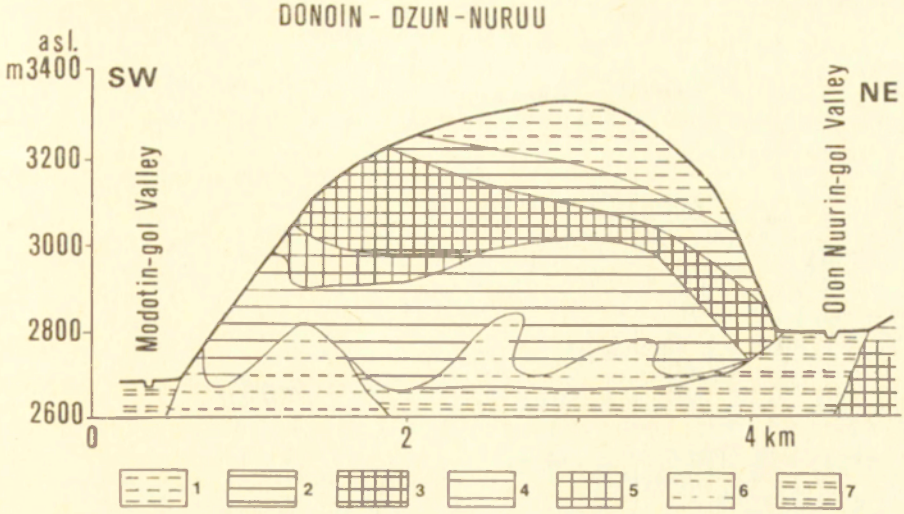


Fig. 20. Altitudinal zones of soils on the Donoin-Dzun-Nuruu ridge

1 — poorly developed primitive, stony, brown soils structure ground underlain by permafrost; 2 — stony brown soils affected by congelifluction, underlain by permafrost; 3 — stony grey-brown soils affected by congelifluction, underlain by permafrost; 4 — hydrogenic chernozems — structure ground underlain by permafrost; 5 — brown chernozems affected by congelifluction, underlain by permafrost; 6 — dark chestnut soils affected by congelifluction, underlain by permafrost; 7 — peaty and peaty-gley soils — structure ground and thufurs underlain by permafrost

shift of climatic belts with time. Because of the varying influences of cryo-humid conditions (from the north) and of cryo-arid conditions (from the south) both typical soils and polygenetic, transitional soils occur at adjacent sites.

Each of soil belts is marked by a definite set of morphogenetic and pedogenic processes. Changes in the hydrothermal conditions related to slope aspect bring about changes in the succession of the altitudinal zones and inversions of asymmetry.

PRIMITIVE BROWN SOILS — STRUCTURE GROUND UNDERLAIN BY PERMAFROST

The mountain tundra belt in which precipitation exceeds evaporation (Dauksza and Soja 1977) extends above 3300 m asl. on south- and east--facing slopes, and above 300 m asl. on north- and northwest-facing slopes. On nearly horizontal surfaces (2—3°) there occur polygonal forms of structure (patterned) ground measuring 0.4—4.0 m across. On steeper surfaces there occur both striped soils and soils of congelifluction lobes. The polygons and sorted stripes which consist of blocks, tabular stones and smaller rock fragments are delineated by furrows, 1—3 m deep, in which snow and ice tend to collect. During the warm season the surplus melt- and rainwaters are freely drained by these depressions. The central areas of 'the polygons have a concentration of fine waste which is

Table 6. Soil belts in the southern Khangai and slope aspect

Soil belt	Donoin-Dzun-Nuruu altitude [m]			Sant valley altitude [m]		
	N	SE	S	N	SE, E	S
Mountain primitive brown soils and brown soils, poorly developed — structure ground underlain by permafrost	> 3040	> 3210	> 3210/3240/	—	—	—
Mountain brown soils — structure ground affected by congeliflunction, underlain by permafrost	2820—3040	2780—3210	3180—3210	2400—2450	—	—
Mountain grey-brown soils — structure ground affected by congeliflunction, underlain by permafrost	2770—2820	2700—3180	2780—2800	—	—	—
Mountain chernozems — structure ground affected by congeliflunction, underlain by permafrost	—	2650—3180	2800—3100	2400—2710	2550—2710	—
Mountain dark chestnut soils underlain by permafrost	—	2600—2980	2680—2700	2150—2500	2700	2550—2650
Mountain light chestnut soils, poorly developed, denuded	—	—	—	—	< 2550	< 2520

a vari-grained fine-sandy loam (tab. 5) of similar mechanical and chemical composition (Kowalkowski 1977). It shows thixotropy. Where meshes of polygons are sod-covered the dispersed clay tends to migrate and to form the Bvt horizon. The low humus content (about 50 t/ha), low total water holding capacity, low capillary water capacity (tab. 6) and the full water saturation of the soil (Kowalkowski 1977) indicate a predominance of both frost action and reduction processes in such soils. Their high acidity (pH 4.0) is due to the removal of bases by the cold water. Tyutyunov (1960), Rakhno (1964) and Stepanov (1975) have emphasized that geochemical processes are activated in a cool soil environment, even when water is frozen, and bacteria increase in number. Soluble humus also occurs in greater amounts there. Chemical processes going on in the montane tundra soils are confirmed by the results of analyses of the quartz grain surface morphology (Kowalkowski and Mycielska-Dowgiałło 1980). The inactive, low-centred polygons are covered with gravels and stones (pavement). The active polygons have high central areas which are cracked into small desiccation polygons.

BROWN SOILS — STRUCTURE GROUND AFFECTED BY CONGELIFLUCTION, UNDERLAIN BY PERMAFROST

Within the altitudinal zone of high-mountain *Kobresia* meadows there occur stony brown soils formed from cryogenic waste products above 2800 m asl. on north- and southeast-facing slopes, and above 3100 m asl. on slopes facing the sun. Islands of soils are found at 2400 m asl. on north-facing slopes which are underlain by shallow permafrost (Kowalkowski and Lomborinchen 1977; Kowalkowski and Pacyna 1977). The characteristic pH 4.4—5.5 indicates weaker acidity of the soils under arid conditions. Brown soils have sorted polygonal forms on the flat surfaces and striped patterns (parallel lines of stones and vegetation-covered stripes) on gradients above 2—3°. Brown soils also occupy cryoplanation terraces with blockfields on the scarps. Such soils occur largely on the margins of tundra soils without swamps.

The cryo-humid brown soils have a thin humus horizon which is disturbed by frost action and congelifluction. The proportions of skeleton show little variation throughout the profile (tab. 5). The Bvt horizon occurs usually beneath the BvA horizon being impoverished in mineral colloids. These soils are less acid (pH 4.4—4.6) than the tundra soils, and they are richer in humus. The water holding capacity of the brown soils which tend to retain capillary moisture is larger (tab. 6). On the active congelifluction lobes with bulging fronts up to 2 m high these soils are freely drained by the stony banks. In places streamlets coming from the tundra have eroded shallow channels.At present these are filled with alluvial gravel, sand and fine sand. On the large stone-banked lobes occurring on the block-covered scarps, and extending more than 100 m downslope the brown soils occupy 20—40% of the lobes. Waste products are actively in motion, and frost action also takes place there.

GREY-BROWN SOILS — STRUCTURE GROUND AFFECTED BY CONGELIFLUCTION, UNDERLAIN BY PERMAFROST

These are transitional between brown soils and mountain chernozems. The cover of grey-brown soils is complete on southeast-facing slopes at altitudes of 2700—3180 m asl. It vanishes on north- and south-facing slopes at altitudes of 2770—2870 m. Under the luxuriantly growing *Kobresia* meadows a thick (to 0.7 m), grey-brown humus horizon is developed. The humus content here is four to five times as great as in tundra soils, and two to three times as great as in brown soils (tabs. 6 and 7). Because of the weaker acidity (pH 4.4—5.1), total water holding capacity and possibility of water replenishment from the numerous streams conditions are favourable for the increased biological activity of the soil and accumulation of humus. The large soil capillary moisture capacity (tab. 7) promoting saturation of the soil, shallow permafrost and gradients of 18—28° facilitate periglacial mass movement. This causes both modification and inversion of the genetic soil horizons (Kowalkowski 1977). Such disturbances may also result from the upfreezing of blocks (Pękala 1975). On convex and sod-covered surfaces the grey-brown soils have numerous stone stripes which occupy 30—40% of the area. In flat sites there is a tendency of surface water to stagnate, and swamps develop. For this reason islands of hydromorphic chernozems and peaty-gley soils occur there.

MOUNTAIN CHERNOZEMS — STRUCTURE GROUND AFFECTED BY CONGELIFLUCTION, UNDERLAIN BY PERMAFROST

In the cryo-humid interior of the Khangai these soils are forming a clear belt extending between 2800 m and 3100 m asl. on south-facing slopes and at 2650—3180 m asl. on southeast-facing slopes. On northern slopes chernozems disappear. Towards the margin of the mountains, where precipitation is equal to evaporation (Dauksza and Soja 1977) the soil cover is complete on northern slopes at altitudes of 2200—3000 m. In the Sant valley, both brown and chestnut chernozems occur at altitudes of 2400—2710 m on northern slopes. Islands of such soils are found on southwest- and west-facing slopes. Toward the south chernozems are associated with the sheltered and moister cryoplanation benches and with congelifluction lobes which occur on northwest- and north-facing slopes at 2700 m asl.

The cryo-humid, non-carbonate chernozem belt coincides with the occurrence of congelifuction lobes. These are either sod-covered (in lower

Table 7. Some soil properties at 0–40 cm within the different soil belts

Soil belt	Profile no.	Humus reserve total [t/ha]	Humus reserve % of total	Total water capacity total [mm]	Total water capacity % of total	Capillary water capacity in % of total water capacity		Water capacity non available for plants in % of total water capacity	
		0–40 cm	0–10 cm	0–40 cm	0–10 cm	0–40 cm	0–10 cm	0–40 cm	0–10 cm
Mountain brown soils, poorly developed – structure ground	037	51.3	26.02	153.8	26.6	49.9	13.7		
Mountain brown soils – structure ground underlain by permafrost	038	79.9	59.70	191.5	27.9	72.0	21.5		
	010	119.9	44.89	188.7	30.1	83.7	27.1	9.52	2.1
Mountain grey-brown soils	039	230.7	34.45	252.8	28.2	70.5	21.4		
Mountain chernozem soils	040	545.1	27.00	245.1	25.0	75.0	19.4	6.23	1.6
	041	617.1	27.28	229.5	25.7	83.2	23.2	8.15	2.2
	009	323.9	30.19	228.6	26.1	82.7	20.6	8.32	2.2
Mountain dark chestnut soils	043	208.6	38.00	234.6	26.3	80.3	20.9	11.8	3.1
	013	242.7	31.80	189.7	28.5	70.2	21.9	10.80	2.8
	014	126.5	39.0	190.4	28.7	66.7	18.6	8.76	2.3
	008	221.5	32.7	194.2	27.9	84.6	24.5	9.40	3.1
Mountain light chestnut soils, poorly developed	006	119.4	29.14	209.4	24.5	74.3	19.0	5.91	1.41
	004	114.1	36.32	199.0	25.1	72.8	20.0	6.15	1.66

and wetter sites) or stony (in steeper and drier sites). The dissolved mineral and organic constituents and their colloidal suspensions are removed by cool water (3—4°C) from the tundra and brown soil belts; they accumulate within the chernozem belt where the water temperature rises to 6—7°C. Thus the soil increase permanently in fertility.

As a consequence, the chernozem soils are very rich in humus (more than 500 t/ha at 0—40 cm depths; tab. 7, prof. 040, 041) and the thickness of the humus horizon exceeds frequently 50 cm. The rather uniform concentration of humus in this horizon being evidenced by the occurrence of 27% of the total humus content at 0—10 cm depths, the high porosity and large capillary water capacity indicate a high biological activity of the chernozems discussed. This is expressed by a well developed crumb structure in spite of pH 3.9—4.7, and by numerous earthworms, since water is available to plants (tab. 7).

Within the larger chernozem patches that are less disturbed by frost action the humus horizon is sharply cut off from the underlying brown waste Bv. This boundary is frequently a stone layer being parallel to the slope surface.

The wetter flats exhibit mosaics of both cryo-hydrogenic chernozems and peaty soils with frequently occurring thufurs. The flood plains of the small streams have peat-like soils and peaty soils. These contain more than 35% of organic carbon and are very porous. Their bulk density is 0.47 g/cm³ (Kowalkowski 1977). In the southern Khangai which at present experiences drier climatic conditions the cryogenic, non-carbonate chernozems show different features. The soil surface here is stabilized by forest which increases humidity and reduces temperatures as compared to the surrounding steppes (Niedźwiedź *et al.* 1975). Near the upper tree line, at 2700 m asl., on the cryoplanation terraces, groups of larches being 350—450 years old lean in different directions thus indicating mass movement. The humus horizon (the reserve of humus here exceeds 320 t/ha; tab. 7) has a pH 4.7—5.4 and a depth of about 40 cm. There is a sharp transition to the brown waste. In flat sites this boundary is a stone layer being parallel to the soil surface. Although the chernozems discussed contain less humus, their total water holding capacity and the capillary water capacity are very large, whilst the water content unavailable for plant growth is low (tab. 7). For this reason the soils discussed are classified as chernozem soils.

According to Stepanov (1975) the non-carbonate mountain chernozems of differing age should be classified into at least two groups; the cryo-humid soils actively developing on the southern slope of the Khangai and the partly relict cryo-arid soils. The latter soils generally correspond to the cryo-arid soils here described and also found in northern and eastern Mongolia (Dorzhgotov 1973); Nogina *et al.* 1977; Undral 1978) marked by a clear slope asymmetry (at 1300—1700 m asl.). However, the cryo-humid chernozems that are being actively formed at the present day have been as yet not mentioned by the above authors. The lack of carbonates is a primary feature of such soils. The appearance of carbonates in the profile indicates a secondary superimposition of cryo-aridity.

DARK CHESTNUT SOILS UNDERLAIN BY PERMAFROST

These occur in areas with arid influences, where evaporation exceeds precipitation (Dauksza and Soja 1977). Dark chestnut soils occupy north--facing slopes at 2000—2700 m asl. They reach to elevations 2700 m on west- and southwest-facing slopes and occur at 2250—2650 m asl. on wetter sites lying on southern slopes. In the valleys that are invaded by dry air masses coming from the south patches of dark chestnut soils are found up to 2900 m asl. (tab. 6, fig. 21). The soil-forming processes really do reflect the prevailing hydrothermal conditions. These soils tend to rejuvenate locally because of truncation of profile due to slope erosion and to deposition of eroded soil materials on the local soils.

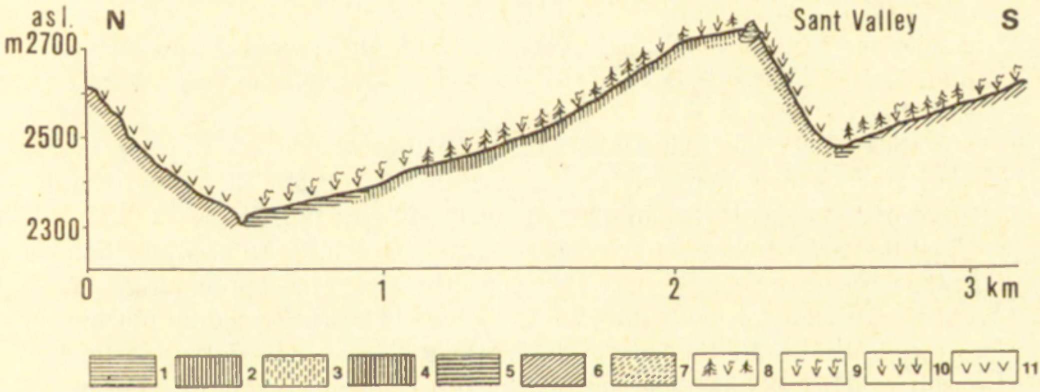


Fig. 21. Soil catenas in the surroundings of the Sant valley (based on data by Haase *et al.* 1964; Kowalkowski and Pacyna 1977)

1 — relict brown soils; *2* — relict brown chernozems; *3* — relict grey chernozems — structure ground; *4* — relict hydrogenic chernozems; *5* — chernozem deluvia; *6* — stony dark chestnut soils; *7* — denuded light chestnut soils; *8* — forest steppe; *9* — mountain steppe; *10* — mountain meadow-steppe; *11* — dry mountain steppe

On the north-facing, lower and mid-slopes the parent materials are thick sequences of stony and gravelly slope sheets (brown waste) overlain by fine-grained material of varying thickness. This includes 22—37% of silt and clay particles. Dark chestnut soils are usually rich in humus to 40 cm depth (125—220 t/ha). The concentration of humus in the profile is irregular, being largest in the upper 10 cm. The acidity is rather high (pH — 6.0); it increases with increasing altitude, but decreases with increasing dryness of climate. The total water holding capacity is as great as that of brown soils, although the capillary moisture capacity is low and variable (tab. 7). This indicates that rainfall sinks readily into these soils. However, the water capacity non available for plant growth (tab. 7) is large. Shallow permafrost occurring on north-facing slopes secures sufficient water for both plant growth and biological activity of soils during the dry seasons. At foot of tors which retreat by frost attack thick dark chestnut soils tend to develop locally (prof. 008, tabs. 6 and 7). These are built up by accumulations of rock debris and dark chestnut soil materials filling the gaps.

POORLY DEVELOPED LIGHT CHESTNUT SOILS

The youngest, azonal cryo-arid soils develop largely on the dry and steep slopes facing the sun, at altitudes below 2550 m (tab. 6). These soils are being formed from the young chestnut waste under arid climatic conditions. The waste contains low amounts of silt and clay particles (11.3—14.4%) and 22—41% of skeleton in the top soil. This is overlain by fine scree produced by present-day exfoliation and thermal disintegration (Pękala 1975; Pękala and Ziętara 1977). Because of the dryness of climate the soils are rather weakly acid (pH 5.2—5.9). On very dry and slightly denuded sites incipient horizons of carbonate precipitation are forming at 20—40 cm depths.

In these soils the humus horizon which developed on the chestnut waste is a few centimetres thick. It contains 2.5—4.5% of humus (total reserve of humuses less than 120 t/ha at 0—40 cm; tab. 7).

The light chestnut soils are characterized by a large total water holding capacity, large capillary water capacity and low water range non available for plant growth (tab. 7). The cause is both the coarse mechanical composition and the accumulational nature of material underlying the soil which itself has been altered by processes of denudation. Because of the scarce cover (less than 25%) of dry, mountain steppe plants (Kowalkowski and Pacyna 1977), marked desiccation and lack of a snow cover in the winter (Froehlich *et al.* 1977) soil conditions are favourable for the occurrence of sheet wash which attacks even flat surfaces. Consequently, in places the soil profile tends to rejuvenate constantly, and both disintegrated rock and slabs of bedrock are exposed to erosion.

ANNA PACYNA

VIII. VEGETATION AND THE PLANT VERTICAL ZONES

In the meridional section of the Khangai Mts there occurs, parallel with the increased aridity of the climate, a transition from a boreal type of vegetation with a well-formed mountain taiga through a breach in the dense forest belt and a decrease in its area towards increasingly drier types of forest and the elimination of the boreal elements until the woodland communities have completely disappeared and been replaced by steppe communities. At the same time, the forest vertical zone advances to increasingly higher altitudes.

The Gobi has not much influence on the flora and plant communities of the southern slopes of the Khangai. Desert steppes do not occur there, nor are there many representatives of Central Asiatic species. In the region studied, *Clematis tangutica* (Maxim.) Korsh. is encountered as well as ***Eurotia ceratoides*** (L.) C.A.M. In the vicinity of the Gobi the xerophytization of the plant communities may be considered only as a result of the increasing aridity of the climate.

In the profile the type of vertical arrangement of the vegetation in the Khangai is also changed. On the northern limits this is a system of boreal type (Karamysheva and Banzragch 1977), characterized by the complete development of vertical zones, with a separate forest vertical zone and well-developed mountain taiga, as well as the subdivision of the high mountain taiga zone into sub-goletz, goletz, and subnival subzones, with communities typical of these sub-goletz open woodlands and mountain tundras, characteristic of the high altitude localities in South and East Siberia (Karamysheva and Banzragch 1977). A similar differentiation of the high-mountain vertical zones is seen in northern Mongolia, in mountains with a boreal type of plant vertical zone — the Khubsugul Mts (Batraeva *et al.* 1976) and the Khentei.

The occurrence of the monutain tundra is associated with the taiga. According to Sochava and Gorodkov (1956) it forms, together with the taiga, one type of vertical zoning and is vertically a substitute for taiga formation. In other parts of the Khangai, which has an extremely continental climate, an arid type of vegetative vertical zones is formed (the

Khangai variant; Yunatov 1950). Karamysheva and Banzragch (1976 b) have termed the zone system on the southern slope of the Khangai the southern Khangai type.

A reduction of some zones, especially the forest zone, is characteristic of the arid system of vegetation zones. The character of the high-mountain zone is also altered. Particular subzones may be eliminated and the prevailing plant communities may be changed. The absence of sub-goletz open woodland and mountain tundra may be connected with the reduction in the vertical zone of mountain taiga. A dominant role is then played by the high-mountain *Kobresia* meadows ("pustoshi" in Russian) typical of the mountains of Central Asia (Karamysheva and Banzragch 1976 b, 1977).

The transition from the boreal to the arid type (the southern Khangai type) of zone arrangement is gradual, depending on slope orientation. The combinations of plant vertical zones also differ in various parts of the Khangai. These were studied by Karamysheva and Banzragch (1977) on a transect carried out along meridian 98.

The comparatively short but steep southern slope of the Khangai, with rapidly decreasing altitudes, varying in relief and climate, and in consequence the conditions of the habitats, is a fruitful field for the investigation of changes in the arrangement of plant vertical zones. The transition from steppes to high-mountain meadows and scree slopes without any closed plant cover may be observed over a relatively small area.

In the Tsagan-Turutuin-gol catchment area [6] the following vertical zones of vegetation may be distinguished:

1) a steppe vertical zone up to 2100 m asl.,

2) a forest-steppe vertical zone (mosaics of mountain steppes and forest [7] 2000/2100—2600/2700 m asl.);

3) a lower high-mountain vertical zone extending above 2600 m asl.,

a) a lower high-mountain vertical subzone, 2600—3000 m,

b) an upper high-mountain vertical subzone over 3000 m.

The arrangement and extent of the plant vertical zones presented coincide as a rule with the landscape zones into which this area has been divided (Haase *et al.* 1964).

THE STEPPE VERTICAL ZONE

A narrow strip of steppes typical of Central Mongolia surrounds the Khangai to the south. The steppes penetrate the bottoms of the wide valleys at a small distance from the mountainous interior. In the Tsagan-

[6] Surveys were carried out in the valleys of the rivers Tsagan-Turutuin-gol, Ulin-gol, Urgotuin-gol, Donoin-gol, Modotin-gol, Olon-nuur and on the Donoin-Dzun-Nuruu ridge as well as the unnamed peak between the Urgotuin-gol and Modotin-gol valleys (3494 m asl.). Detailed studies were made in the Sant valley.

[7] in Yunatov's sense.

-Turutuin-gol valley, *Artemisia frigida—Stipa Krylovii* steppe occupies the higher terraces with light chestnut soils (Kowalkowski and Lomborinchen (1975). The type of steppe described is presented by two phyto-sociological records[8] (tab. 8, records 5, 6). Record 6 illustrates the early summer aspect with flowering *Astragalus galactites* Pall. and not yet fully grown grasses. Record 5 was made in the height of summer, when the grasses had already ears, while *Astragalus galactites* Pall. had finished flowering, its herbage had dried, and its share in the community had become less evident. The *Artemisia frigida—Stipa Krylovii* steppe is a community with a rather sparse low herbage, consisting entirely of xerophytes. The species composing this community are characteristic of the Mongolia steppe plains: *Koeleria cristata* (L.) Pers., *Poa attenuata* Trin. and *Stipa Krylovii* Roshev. *Artemisia frigida* Willd. is an important species among the dicotyledonous plants. Because of the great altitude, species typical of the mountain steppes are also found here; *Arenaria capillaris* Poir. and *Festuca lenesis*. Drob. appear in abundance.

THE FOREST-STEPPE VERTICAL ZONE

The forest-steppe vertical zone occupies intermediate mountain situations in which predominance of mountain steppes is characteristic. Among these, small forest patches are scattered on north-facing slopes. Such a distribution of the plant cover is due to slope asymmetry in the valleys running from east to west. The effect are the extremely different habitat conditions (Kowalkowski and Pacyna 1977) prevailing on north- and south-facing slopes. This is caused by differences in the thermal and water balances, in relief and slope processes, and in the soil cover (Niedźwiedź *et al.* 1975; Starkel 1975; Pękala 1975; Kowalkowski and Lomborinchen 1975; Pękala and Ziętara 1977).

In the catchment area of the Tsagan-Turutuin-gol woodlands occupy only a small total area (tab. 3), appear only on N, NNW, and NNE-facing slopes, and are often restricted to concave slopes. These slopes, cooler and wetter (Niedźwiedź *et al.* 1975), form the most favourable habitat for forest in these climatic conditions. The occurrence of permafrost on them is an important factor (Kowalkowski and Lomborinchen 1975), as it decides the mode of the water supply in conditions of air humidity deficit (Bannikova and Khudyakov 1977). It is also the cause of cryogenic processes reflected in the habit of the forest ("drunken trees").

While mountain larch and *Larix sibirica—Pinus sibirica* mountain taigas, fairly well-developed in places, occur in the northern Khangai (Yunatov 1950; Korotkov 1976; Karamysheva and Banzragch 1977), larch

[8] The Braun—Blanquet method has been used for the phyto-socialogical records.

Table 8. Steppes and mountain steppes

Number of record in table	1	2	3	4	5	6	7	8	9	10
Field number of record	49	36	37	73	61	22	64a	71	93	43
Date	23.7	10.7	10.7	6.8	28.7	30.6	30.7	1.8	15.8	19.7
Altitude [m]	2400	2520	2440	2450	2075	2055	2410	2420	2420	2480—2500
Slope aspect	NE	NEE	NEE	W	W	SW	NWW	SSW	SSE	S
Inclination	5	20	5	6—10	2	1—2	2	5—10	40	20
Coverage of herb layer (C) [%]	95	90	95	85	85	80	90	50	50	50
Coverage of mose and lichen layer (D) [%]	10	0	10	1	1	1—2	0	0	0	0
Area of record in sq. m	200	300	...	300	2000	100	400	300	400	200
1	2	3	4	5	6	7	8	9	10	11
Koeleria cristata (L.) Pers.	+	2.2	3.3	3.2	3.2	4.3	3.2	2.2	1.2	+2
Poa attenuata Trin.	2.2	+	.	3.2	1.2	+	3.2	2.2	1.2	3.2
Festuca lenensis Drob.	.	4.2	3.3	3.2	+2	2.2	2.2	2.2	2.2	.
Agropyron cristatum (L.) Gaertn.	.	+2	+	+	1.2	1.2	2.2	+	+	.
Artemisia pycnorhiza Ldb.	.	1.1	3.2	3.2	+	+	1.2	2.2	+	.
Amblynotus obovatus (Ldb.) I. Johnst.	.	+	+	+	+	+	+	1.1	1.1	.
Dontostemon integrifolius (L.) C.A.M.	.	+	+	.	1.1	+	+	+	+	1.1
Thalictrum petaloideum L.	+	2.1	.	1.2	.	.	2.2	2.2	+	1.1
Iris flavissima Pall.	.	1.2	.	1.2	+	+	+	+	+	+2
Potentilla bifurca L.	.	.	+	1.2	1.2	+	1.2	1.1	.	+
Pulsatilla ambigua (Turcz.) Juz.	1.2	2.2	.	3.2	.	.	+	+	.	+
Androsace septentrionalis L.	+	+	.	+	+	.	+	+	.	.
Pulsatilla sp.	.	+	2.2	2.2	.	.	1.2	2.2	+	.
Potentilla cfr. *sericea* L.	.	.	.	+	2.2	2.2	2.2	+	+	.
Echinops dahuricus Fisch.	.	.	.	+	+	.	+	+2	2.2	+
Androsace incana Lam.	1.2	+	+	+2	.	+	.	.	.	.
Leontopodium ochroleucum Beauv. s. l.	1.1	+	2.2	2.2	.	.	+	.	.	.
Gentiana decumbens L. f.	+	+	+	1.2	.	.	1.2	.	+	.
Sanguisorba officinalis L.	2.2	1.2	.	+	.	.	+	.	.	2.2

1	2	3	4	5	6	7	8	9	10	11
Veronica ciliata Fisch.	+	+	.	.	.	.	+	.	+	+
Oxytropis filiformis DC.	.	+	2.2	+	+2	1.2	.	.	.	.
Polygonum angustifolium Pall.	.	+	.	+	.	.	.	+	+	1.2
Rheum sp.	.	+	+	.	.	.	+2	+2	.	+
Carex duriuscula C.A.M.	.	.	+2	.	+2	+	1.2	+	.	.
Rhodiola rosea L.	+	+	.	.	.	.	.	+	.	1.2
Helictotrichon Schellianum (Hack.) Kitag.	1.2	1.2	.	1.2	.	.	.	.	.	.
Dasiphora fruticosa (L.) Rydb.	.	+	.	.	.	.	+2	.	+	2.2
Sibbaldianthe adpressa (Bge.) Juz.	.	+	.	.	1.2	+	.	1.2	.	.
Arenaria capillaris Poir.	.	.	+	.	3.2	2.2	.	.	+	.
Thalictrum foetidum L.	.	.	.	2.2	.	.	1.2	.	1.2	1.1
Oxytropis nitens Turcz.	.	.	.	+2	.	.	.	+	+2	2.3
Euphorbia discolor Ldb.	.	.	.	+	.	.	2.2	+	.	+
Pedicularis abrotanifolia M.B.	.	.	.	+	.	.	.	1.1	+	+
Stipa Krylovii Roshev.	.	.	.	.	1.2	+2	.	+2	+	.
Aster alpinus L.	+	+	.	+	.	.	.	.	.	.
Oxytropis strobilacea Bge.	+	+	+	.	.	.	.	.	.	.
Senecio campester (Retz.) DC.	+	+	+	.	.	.	.	.	.	.
Allium prostratum Trev.	.	+	+	.	1.2	.	.	.	.	.
Artemisia changaica Krasch.	.	.	.	+	+	.	.	+	.	+
Thermopsis lanceolata R. Br.	.	.	.	+	.	.	.	1.1	.	+
Linum baicalense Juz.	.	.	.	+	.	.	.	.	+	+
Heteropappus altaicus (Willd.) Novopokr.	.	.	.	+	+	.	.	.	+	.
Orostachys spinosa (L.) C.A.M.	.	.	.	.	2.2	+	.	.	+	.
Chamaerhodos erecta (L.) Bge.	.	.	.	.	+	.	.	+	.	+
Limonium flexuosum (L.) Ktze.	.	.	.	.	.	.	+2	+	.	+
Thymus gobicus Tscherm.	.	.	.	.	.	.	.	+	+	+
Taraxacum sp.	+	+	+	.	.	.	.	.	.	.
Umbelliferae indet.	.	.	+	+	.	.	+	.	.	.
Artemisia frigida Willd.	.	.	.	.	3.2	3.2	.	.	.	.
Galium verum L.	2.1	+	.	.	.	.	.	.	.	.
Pedicularis rubens Steph.	2.2	+	.	.	.	.	.	.	.	.

1	2	3	4	5	6	7	8	9	10	11
Peucedanum hystrix Bge.	.	.	.	.	.	.	.	2.2	+	.
Astragalus galactites Pall.	.	.	.	.	+3	1.3	.	.	.	.
Cotoneaster melanocarpa Lodd.	.	.	.	.	.	.	.	.	1.3	1.3
Carex Korshinskyi Kom.	.	.	.	+2	.	.	.	1.1	.	.
Pedicularis achilleifolia Steph.	.	+	.	.	.	.	.	.	1.2	.
Zerna pumpelliana (Scribn.) Tzvel.	+2	+2	.	.	.	.	.	.	.	.
Bupleurum sp.	.	.	.	.	+2	.	.	+	.	.
Gentiana pseudoaquatica Kusn.	.	.	.	+	.	.	+	.	.	.
Plantago depressa Willd.	+	+	.	.	.	.	.	.	.	.
Kobresia Bellardii (All.) Degl.	+	+	.	.	.	.	.	.	.	.
Melandrium brachypetalum (Hornem.) Fenzl.	+	+	.	.	.	.	.	.	.	.
Stellaria dichotoma L.	.	.	.	.	.	.	.	.	+	+
D										
Mosses undetermined	+	.	.	.	.	.	.	.	.	.
Lichens										
Parmelia vagans Nyl.	.	.	+	+	+	+	.	.	.	.

Record 1[1]. *Festuca Kryloviana* Reverd. 4. 3, *Silene repens* Patr., *Achillea asiatica* Serg., *Dendranthema Zawadzkii* (Herb.) Tzvel. *Cerastium arvense* L., *Carex obtusata* Liljebl., *Dianthus versicolor* Fisch., *Polygonum viviparum* L. 2.1, *Draba nemorosa* L., *Trisetum sibiricum* Rupr., *Mertensia ochroleuca* k.-Gal., *Gentiana macrophylla* Pall., *Myosotis asiatica* Schischk. et Serg., *Artemisia* sp., *Oxytropis* sp. Record 2. *Eritrichium rupestre* Bge., *Androsace Bungeana* Schischk. et Bobr., *Delphinium dissectum* Huth, *Stellaria cherleriae* (Fisch.) Williams., *Anemone crinita* Juz., *Hedysarum inundatum* Turcz., *Polygonum alpinum* All., *Potentilla* sp. 1.2. Record 3. *Erysimum altaicum* C.A.M. Record 4. *Veronica incana* L. 1.2, *Pedicularis myriophylla* Pall., *Linaria acutiloba* Fisch. Record 5. *Potentilla acaulis* L., *Artemisia palustris* L., *Thesium longifolium* Turcz., *Oxytropis selengensis* Bge. Record 6. *Poa* cfr. *pruinosa* Korotky 1.2, *Caragana pygmaea* (L.) DC., *Oxytropis* sp. Record 7. *Leymus secalinus* (Georgi) Tzvel. 1.2, *Carex melananthaeformis* Litw., *Oxytropis* sp. Record 9. *Chamaerhodos altaica* (Laxm.) Bge. 1.2, *Oxytropis tragacanthoides* Fisch., *Ptilotrichum tenuifolium* (Steph.) C.A.M., *Allium leucocephalum* Turcz., *Serratula centauroides* L., *Silene jenisseensis* Willd. Record 10. *Valeriana officinalis*. K. 2.2, *Artemisia santolinifolia* Turcz. 1.2, *Potentilla viscosa* G. Don. 2.2, *Scutellaria scordiifolia* Fisch., *Allium altaicum* Pall., *Artemisia monostachya* Bge., *Melandrium apricum* (Turcz.) Rohrb., *Vicia megalotropis* Ldb., *Atragene sibirica* L.,

Localities: 1 — Ulin-gol valley, 2,3 — Donoin-gol valley, 4,7,8,9,10 — Sant valley, 5,6 — Tsagan-Turutuin-gol-valley

[1] The symbol "+" has been omitted.

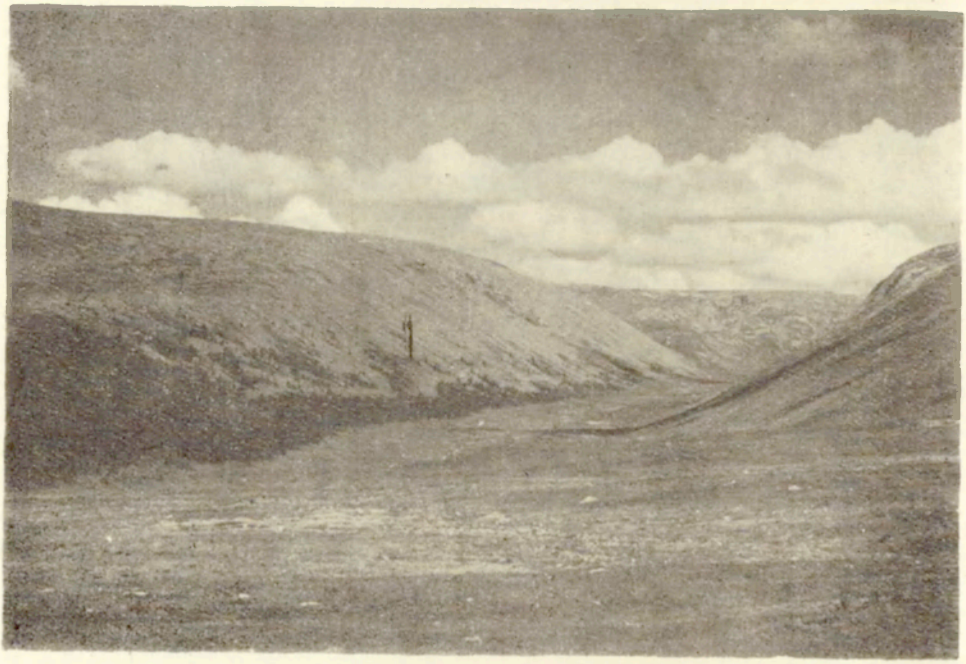
forests defined by Korotkov (1976) as pseudo-taiga forests grow in the southern Khangai. In the extremely arid conditions at the southern limit of the boreal type of forest of the eastern Siberian mountain taiga there remains only *Larix sibirica* Ldb. *Pinus sibirica* (Rupr.) Mayr requires more moisture than larch (Sochava 1956). The boreal taiga species of undergrowth and herb layer have been eliminated, with the exception of *Atragene sibirica* L. In other parts of the Khangai there still occur *Linnaea borealis* L., *Pyrola incarnata* Fisch. and *Vaccinium vitis-idaea* L. (Karamysheva and Banzragch 1976 a, 1977). The type of forest in equation is represented in the phytosociological records shown in table 9.

Although the individual patches of the pseudo-taiga type of forest resemble one another fairly closely (Korotkov 1976), a certain differentiation depending on the conditions of humidity appears in the Tsagan--Turutuin-gol catchment area. The patches situated nearer the southern borders of the Khangai, exposed to the action of the hot dry southern winds (Avirmid *et al.* 1976), differ somewhat in profile from those situated in the interior of the mountains, where the atmospheric precipitation is higher (*vide* chapter IV). There is a difference in the height of the trees and the luxuriance of the herb layer. Tree stands on dry slopes are often dispersed, sometimes assuming a park-like character.

The forest stand on the southern slopes of the Khangai is rather low (15—20 m on the average). There are no other trees besides *Larix sibirica* Ldb. The undergrowth consists entirely of young larches. The forest patches appearing on the blockfields, where the canopy cover is less dense, form an exception, for here numerous shrubs occur: *Lonicera altaica* Pall., *Ribes altissimum* Turcz., *Rosa acicularis* Lindl. and Berberis sibirica Pall. (tab. 9, record 5).

The herb layer is formed by mountain steppe species (*cf.* tab. 8, records 1, 2). Among these may be distinguished the meadow-woodland mesophytes *Trollius asiaticus* L., *Geranium pseudosibiricum* J. Meyer, *Poa sibirica* Roshev., *Sanguisorba officinalis* L. and *Anemone crinita* Juz. The meadow-steppe species include *Aster alpinus* L., *Cerastium arvense* L., *Artemisia tanacetifolia* L., *Galium verum* L. and *Zerna pumpelliana* (Scribn.) Tzvel. The high mountain species include *Polygonum viviparum* L., *Pedicularis verticillata* L., *Ptilagrostis mongolica* (Turcz.) Griseb., *Festuca altaica* Trin., *Viola biflora* L., and *Hedysarum inundatum* Turcz. The latter occur more frequently in the higher situated forest patches (tab. 9, record 5), owing to the contact of the forests with the high mountain meadows. The most frequent mosses are the xerophytes *Rhytidium rugosum* (Hedw.) Kindb. and *Abietinella histricosa* (Mitt.) Broth.

The extremely continental climate causes the forest patches in the Khangai to rise to considerable heights. In connection with this there is not only an upper but a lower forest limit. The woods on the southern slopes of the Khangai lie higher than in the other parts.



Phot. by A. Pacyna

Phot. 10. The formerly glaciated Modotin-gol valley with a larch forest patch reaching its upper limit in the left foreground the mountain steppes. The representatives of *Artemisia* genus are absent or appear only in small numbers, except for *Artemisia tanacetifolia* L.

Drier variants of the mountain steppes occupy the rest of the slope and large areas in the valley bottoms. They are marked by the lesser density and lower growth of the herbage. While the mountain meadow steppes are associated with mountain chernozems. other variants appear on the chestnut soils.

Records 3 and 4 (tab. 8) present a variant of *Leontopodium ochroleucum* Beauv. s.l., and record 7 shows a variant of *Agropyron cristatum* (L.) Gaertn., *Gentiana decumbens* L. f., and *Thalictrum petaloideum* L.

The stony mountain steppes (tab. 8, records 9, 10) being associated with south-facing talus slopes are extrazonal communities, entering into the vegetation of regions situated further south (Yunatov 1950). On account of the thermal and humidity conditions prevailing there (Niedźwiedź *et al.* 1975) and the nature of the waste covers and slope processes (Starkel 1975; Pękala 1975; Pękala and Ziętara 1977; Kowalkowski and Lomborinchen 1975) these slopes are unfavourable for vegetation habitat. Xerophytic grasses predominate among the markedly dispersed vegetation. In some stands numerous bushes appear, including *Dasiphora fructicosa* (L.) Rydb., *Cotoneaster melanocarpa* Lodd. and *Spiraea flexuosa* Fisch. Tall forbs with greater moisture requirements, e.g. *Sanguisorba officinalis* L. and *Valeriana officinalis* L., take advantage of the more humid micro-habitats amid the boulders. The deluvial sheets at the bases of slopes with a southern aspect are the driest environments. A very dry variant of mountain steppe with dispersed low vegetation (tab. 8, record 8) overgrows soil covered by a thick layer of slope derived material. Xeromorphic grasses (*Poa attenuata* Trin., *Koeleria cristata* (L.) Pers. and *Festuca lenensis* Drob.) predominate. *Artemisia pycnorhiza* Ldb. appears in plenty.

Only the most important communities of the forest-steppe vertical zone have been presented. The meadows sometimes occurring on the valley bottoms and occupying a small area in relation to the whole vertical zone are discussed in another paper on the Sant valley.

THE HIGH-MOUNTAIN VERTICAL ZONE

The high-mountain vertical zone in the Tsagan-Turutuin-gol catchment is subdivided into the lower and the upper subzones within the arid type of vertical plant zones.

THE LOWER HIGH-MOUNTAIN VERTICAL SUBZONE

In the lower high-mountain vertical subzone the vegetation on the slopes differs from that in the valley bottoms.

High-mountain *Kobresia* meadows are predominant on the slopes and are characterized by the following phytosociological record.

Record no. 74, 9 VII 1975, slope of the Donoin-Dzun-Nuruu above the Olon-nuur valley, 2770 m asl., east-facing slope, inclination 10°, area of record 400 m^2, herb layer cover 100%, ground layer cover 50%, height of herbage up to 15 cm, height of inflorescences of *Polygonum alopecuroides* and *Sanguisorba officinalis* up to 30 cm. *Kobresia Bellardii* (All.) Degl. 4.3, *Sanguisorba officinalis* L. 3.2, *Poa attenuata* Trin. 2.2, *Festuca lenensis* Drob. + *F. ovina* L. 1.2, *Polygonum viviparum* L. 1.2, *Campanula Turczaninovii* Fed. 1.2, *Potentilla nivea* L. 1.2, *Crepis polytricha* Turcz. 1.2, *Ptilagrostis mongolica* (Turcz.) Griseb. 1.2, *Dasiphora fructicosa* (L.) Rydb. 1.3, *Helictorichon altaicum* Tzvel. 1.2, *Senecio campester* (Retz.) DC. 1.2, *Polygonum alopecuroides* Turcz.[9], *Pulsatilla ambigua* (Turcz.) Juz., *Anemone crinita* Juz., *Leontopodium ochroleucum* Beauv. s.l., *Oxytropis strobilacea* Bge., *Artemisia tanacetifolia* L., *Carex macrogyna* Turcz., *Poa sibirica* Roshev., *Koeleria cristata* (L.) Pers., *Androsace Bungeana* Schischk. et Bobr., *Polygonum angustifolium* Pall., *Pedicularis Oederi* Vahl, *Thalictrum alpinum* L., *Thlaspi cochleariforme* DC., *Arenaria capillaris* Poir., *Aster alpinus* L., *Rhodiola rosea* L., *Stellaria petraea* Bge., *Rumex acetosa* L., *Melandrium brachypetalum* (Hornem.) Fenzl., *Allium strictum* Schrad., *Gentiana macrophylla* Pall., *Elymus confusus* (Roshev.) Tzvel. Mosses: *Aulacomnium palustre* var. *imbricatum* B.S.C. 3.3. Lichens: 1.2 (*Cladonia pyxidata* (L.) Fr., *Parmelia vagans* Nyl., *Pertusaria* sp.).

There is a certain proportion of steppe species in the *Kobresia* high mountain meadows, also observed by Yunatov (1950). This has been caused on the one hand by the aridity of the climate and on the other by direct contact with the mountain steppes on the treeless slopes. *Festuca lenensis* Drob., characteristic of the mountain steppes, appears fairly abundantly besides the steppe grasses *Poa attenuata* Trin. and *Koeleria cristata* (L.) Pers. Other species typical of the mountain steppes are *Artemisia tanacetifolia* L., *Polygonum angustifolium* Pall. and *Arenaria capillaris* Poir. A group of high mountain species is also fairly abundant: *Kobresia Bellardii* (All.) Degl., the predominant species in this community, *Polygonum viviparum* L., and *Thalictrum alpinum* L., mentioned by Yunatov (1950) as a subdominant of the *Kobresia* high-mountain meadows, as well as *Carex macrogyna* Turcz., *Crepis polytricha* Turcz., *Androsace Bungeana* Schischk. et Bobr., *Ptilagrostis mongolica* (Griseb.) Turcz., and *Pedicularis Oederi* Vahl.

These meadows reach an altitude of about 2900 (3000) m asl. At their upper limit the dense plant cover is interspersed by stony lobes. The higher it is, the greater the area occupied by rocks, and the *Kobresia* meadows shrink to small stands between these.

The lower limit of the high-mountain vertical subzone in the valley bottoms is decreased to ca 2500 m as the result of the influx of colder air masses. As a rule the flat valley bottoms on the main Khangai ridge are well covered with peat. This has been caused by the accumulation

[9] The sign + has been omitted.

of both rainwater and water from the thawing permafrost. In addition the valley bottoms are fed by waters from the slopes and extensive plateaus on the mountain ridges and peaks, where precipitation amounts are higher than in the valleys (vide chapter VI), and snow-patches melt in summer.

Marshy hummocky high-mountain meadows are formed on the peaty substrate, where sedges predominate (including *Carex microglochin* Whlbg., *C. dichroa* Freyn, and *C. norvegica* Retz.), *Kobresia sibirica* Turcz., *Allium schoenoprassum* L., *Liguria sibirica* (L.) Cass. and *Eriophorum Scheuchzeri* Hoppe are also frequent. The proportion of *Saxifraga hirculus* L., *Koenigia islandica* L. and *Claytonia Joanneana* R. et S. is characteristic. Mosses are abundant, including *Philonotis fontana* (Hedw.) Brid. and *Compylopus Schimpferi* Milde.

Above altitudes of about 2800 m, as a result of the increasing slope inclination, the substrate is less waterlogged, and high-mountain meadows develop with *Trollius asiaticus* L., and willow thickets appear including *Salix berberifolia* Pall.

THE UPPER HIGH-MOUNTAIN VERTICAL SUBZONE

Areas extending above 3000 m may be considered as the upper high--mountain subzone. The slopes bear huge blocks and smaller boulders. There is no longer a dense plant cover. Single tufts of vascular plants, mosses and lichens exist between the boulders. The cushions of turfy cryopetrophytes are increasingly important among the scanty vegetation.

Stony mountain tundra occupies the broad flat cryoplanation terraces on the ridges and mountain tops. On an unnamed peak (3494 m asl.) occurring between the valleys the Modotin-gol and Urgotuin-gol, polygons developed as a result of frost segregation.

Their central areas are covered with fine materials and almost bare of vegetation. Vegetation appears abundantly in the frost cracks bordering the polygons, where the vital conditions are more suitable. Vegetation here finds some shelter from the strong winds which injure the plants mechanically and the same time increase evaporation, with a total humidity deficit. The lack of fines, which would retain water through the soil colloids is the cause of the escape of both the rainwater and the water derived from the thawing snow and permafrost, and subsequently the subsurface flow down the slopes.

In the frost cracks sterile specimens of *Lagotis integrifolia* (Willd.) Schischk. appear very abundantly.

A somewhat different type of stony mountain tundra appears on the Donoin-Dzun-Nuruu ridge, where stones are bordering the polygons and the central areas bear primitive brown soils (Kowalkowski 1977), there grows a sparse vegetation.

The vascular plant flora in the upper high-mountain vertical zone is very scanty (ca. 25 species), and is composed of two groups of plant: arctic-high-mountain species, some of which are restricted to the Siberian Arctic and Siberian mountains, e.g. *Arenaria formosa* Fisch., while others have a wider distribution, e.g. *Melandrium apelatum* (L.) Fenzl., *Oxygraphis glacialis* (Fisch.) Bge., and *Eritrichium villosum* (Ldb.) Bge. The group of Siberian high-mountain species in considerably more numerous, e.g., *Lagotis integrifolia* (Willd.) Schischk., *Draba ochroleuca* Bge., *Ranunculus altaicus* Laxm., and *Saussurea* cfr. *Pricei Simps.* Mosses and lichens are abundantly represented.

The stony mountain tundra on the central ridge of the Khangai (Brzeźniak 1977) resembles climatically and morphologically the arctic stony mountain tundra (Balandin 1978), but the flora is markedly different.

The stony tundra on the peak 3494 m asl. resembles the "spotty" tundra occurring at the northern limits of the tundra zone, which forms the transition to the arctic deserts (Sochava and Gorodkov 1956).

The upper high-mountain vertical zone in Central part of the Khangai forms a strong contrast with the vegetation in lower vertical zones and the steppes of the southern foothills. It belongs to the regions of the taiga and even tundra zones. The transitory character of the Khangai is also revealed in this (Yurtzev 1977).

LESZEK STARKEL

IX. ALTITUDINAL ZONES IN MOUNTAINS WITH CONTINENTAL CLIMATES

Continental climates exhibit high annual and diurnal temperature variations, low precipitation amounts and a moisture deficit. In any zone which experiences continentality of climate a mountain climate will be similar to climate primarily controlled by latitude, but there are important differences. In low latitudes mountains have stronger diurnal than seasonal temperature changes, while in higher latitudes large annual temperature amplitudes are dominant (Troll 1962). Large-scale variations in both annual and diurnal temperatures tend to occur in the temperate latitude mountains having a continental climate. In any zone the precipitation régimes in the mountains are controlled by latitude (Grigorev and Budyko 1956). The existing altitudinal zones are determined by the limiting factors to the height-dependent thermal and moisture conditions which promote certain physical, chemical and biological processes.

To these factors the zonally arranged slope catenas are added. These are associated with the partially inherited arrangement of landforms (Krivolucky 1971) and governed by the gravitation induced circulation of energy on the mountain slopes (Höllermann 1975). In the mountains, there are said to be either catenas of erosional type marked by a differential course of degradiation along the slope or catenas of hydromorphic type dominated by the effects of impeded drainage (Ollier 1976).

The belt-like arrangement of both climatic and soil-topographic relationships is further complicated by the varied lithologic-tectonic conditions, asymmetry and inversions. If there is not normally enough water in the mountains, then the presence of water appears to affect strongly the existing zones (Messerli 1973).

THE THERMAL BELTS

Attempts were made to define the altitudinal zones of climate by taking temperatures as the criterion (Hess 1965). It appears, however, that temperatures tend to change with height in a rather uniform manner.

It is the various interactions of geoecological factors that determine the existence of a clearly defined boundary. Many phenomena controlled by temperature, and temperature falls of as much as 0.9°C with each 100 m increase in height have been recorded in the mountains with arid climates (Messerli 1973). Annual temperature ranges also decrease with height (reduced continentality), e.g. from 90°C at the foot of the Western Altai Mts to 23°C on the tops of the range (Chelpanova 1963). For the line at which not enough heat is available to melt the winter snow cover the term snow line is used. When the heat balance of the ground surface is negative (mean annual temperatures below −1°C or −2°C), the permafrost zone is produced. The amount of heat received in the summer (growing season) is reflected in the position of the upper boundary of both forests and other plant associations, in the accumulation of organic matter in the soil etc. On the other hand, too high summer temperatures place lower limits on the development of ecosystems because of disturbance of the water balance. Since long-continued inversions of temperature are a frequent phenomenon in the mountains with a continental climate, the geoecological belts tend to be inverted in the inland basins; the tundra belt in the mountain-enclosed basins of Siberia is a good example (Kalesnik 1973).

THE HYDRIC BELTS

Since mountains pick up moisture, the precipitation totals tend to increase with height below an inversion on the tops (Weischet 1969). At the same time evaporation is reduced with height, and so does the water deficit in the mountains which have a dry climate. This deficit being called aridity margin (Troll 1962) is responsible for the position of the lower limit of forest associations. In the Tibesti Mts which experience an extremely dry climate one can hardly speak of any snow line and upper tree line at all (Messerli 1973).

On slopes the distribution of water is partially controlled by runoff, and for this reason the foot-slopes may be abnormally moist. This leads to the salinity increase under arid conditions. On the steep, scree-covered mountain slopes drainage is rapid. Consequently, suitable conditions for local water storage exist in flat-lying areas (summit and slope flattenings, foot-slopes).

THE PERMAFROST BELTS

In the mountains having a continental climate the permafrost becomes more continuous with height because of intense winter cold and short duration of the winter snow cover (Gravis 1974). Similarly, the accompanying active layer decreases in depth with height. Under arid conditions permafrost indicates both the negative heat balance of the ground and the availability of water. It is necessary to distinguish clearly between "dry" frozen and "wet" frozen ground (ice lenses and ice wedges). Van Everdingen (1976) applies the terms "cryotic" (soil temperatures below 0°C) and "frozen" (soil becoming cemented by frozen water) to these phenomena. In mountains with moisture limitations permafrost is often found only on northern slopes. Permafrost is better developed in flat sites where it favours cryoplanation processes (Sukhodrovsky 1975).

THE GEOECOLOGICAL ZONES

The geoecological zones reflect complex thermal-hydric conditions prevailing in the vertical profile. There are often said to be vertical zones of vegetation. These closely affect those of the soil cover which may be more or less complete. Consequently, various morphogenetic processes may develop. The tree line together with the vertical extent of various plant species is mostly restricted by the possibilities of photosynthetic productivity of plants (Tranquillini 1966). In a particular mountain range the vegetation belts depend on the general climatic zone and on the history of vegetation in this area (Karamysheva, Banzragch 1977), so that a large variety of both humid and arid plant and soil belts can be found in Central Asia. The soil changes with height also depend on the vertical exchange and transfer of both water and mineral components. For this reason, the soil may reflect not just the altitudinal zone of climatic but the mountain slope catena. It is possible to recognize two distinct morphogenetic zones above the upper tree line. In the lower zone congelifluction processes are dominant, and the extent of constricted congelifluction often corresponds to that of the alpine meadow zone. The higher scree zone with frost segregation phenomena reaches up to the snow line. As continentality increases the scree zone tends to extend (Hövermann 1962).

SLOPE ASYMMETRY

Asymmetry can be considered in terms of whole mountain chains with marked contrasts in radiation balances and in precipitation amounts (Troll 1972; Weischet 1969). Asymmetry also is found in the particular mountain valley. According to Flohn (1974), in the Turkestan mountains with slopes of 31—33°, net radiation varied between 145 kcal/day on a nothern slope and 426 kcal/day on the opposite slope. In the arid zone cryogenic processes may reach as much as 1500 m above the climatic snow line on the southern mountain sides, whereas on the northern sides glaciers may descend 1000 m below the snow line (Klaer 1969). In Mongolia the occurrence of both intense evaporation and rhythmical daily temperature variations is the basic characteristic of the southern slopes. On north-facing slopes conditions are favourable for the occurrence of patches of permafrost at 2000 m asl., i.e. 600 m lower than on the southern slope (Richter *et al.* 1963).

THE PALEOGEOGRAPHIC ASPECTS OF ALTITUDINAL ZONES

During the Quaternary migrations of the different geoecological belts took place. Indications are the traces of valley glaciations (Selivanov 1972), inactive cryoplanation terraces (Gravis 1974; Kowalkowski *et al.* 1977) and a few pollen diagrams from Mongolia (Vipper *et al.* 1976). An approach to the problem of sequences of relief, soil and plant changes is to study the effects of thermal and moisture changes under continental conditions (Ravsky 1972), where forests are forming only a narrow band on the mountain-sides (fig. 22). Under such conditions a climatic warming would cause the snow line to move upwards, together with the upper and lower tree lines. The effects of moisture changes would be different (Starkel 1977). Increased humidity would cause the height of both snow line and lower tree line to be diminished and that of the upper tree line to be increased. At the same time the forest zone would extend over drier habitats. Changes would occur in the opposite direction with increased aridity of climate. The effects of increased humidity might be expected to be similar to those of cooling causing the southward extension of permafrost. For instance, the relict appearance of *Larix sibirica* in the Gobi Altai Mts may be explained by the former existence of a permafrost bridge which promoted migrations.

THE MOUNTAIN PECULIARITIES IN THE CONTINENTAL CLIMATE OF ASIA

To understand the climate of interior Asia it is necessary to look at the complexity of causes of the marked northward extension of semi--deserts. At the same time the permafrost reaches far to the south and so does the narrow belt of larch forests.

The snow line, together with the upper (thermal) and lower (hydric) tree lines, is rising in height towards the equator (fig. 22).

The prime factor is the radiation balance. There is an increase in both mean annual temperatures and temperatures of the growing season towards the equator. Because of the low air humidity the thermal gradient per 1° of latitude is high. The upper tree line and the snow line are rising in height towards the south due to thermal influences (comp. Hollermann 1973).

Secondly, both precipitation and humidity decrease rapidly towards the high-pressure area in the subtropics. Thus the boundaries of the altitudinal zones further rise in height towards the Tibet where the upper

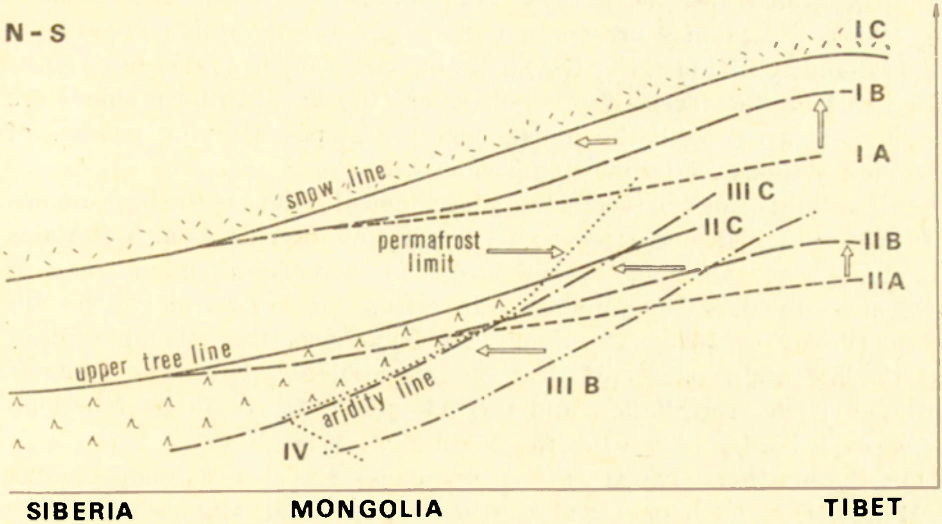


Fig. 22. Complex controls of physico-geographical belts in the mountains of Central Asia which experience continentality of climate

I — position of the snow line due to: IA — height-controlled changes in temperature; IB — dryness increasing equatorwards; IC — shift in the dry zone in the mountain-enclosed continental interior. II — position of the upper tree line due to IIA, IIB and IIC (for explanation see IA—IC). III — position of the aridity margin (lower tree line) due to: IIIB — dryness increasing equatorwards; IIIC — poleward shifting of the aridity margin in the dry continental interior. IV — permafrost limit — inverted within the inland basins and advancing southward as humidity increases at higher elevations in the mountains. Tree-like symbols indicate the vanishing forest belt

tree line extends as high as 4600 m asl. and the snow line reaches a peak value of 6400 m asl. In the mountain ranges that surround in the north the Tibetan Plateau and also occur in its interior the upland steppe comes in contact with the high-mountain tundra, and even the steppe belt is missing there (Murzayev 1966). This also is because of the low moisture content in the high-mountain air (von Wissman 1960).

A third factor producing the climatic-geoecological belts in the interior of Asia are the large dimensions of the continent. In the temperature latitudes moisture influx to Central Asia is caused by a flux from the distant Atlantic (Dzerdzeyevsky 1963). The ephemeral snow cover favours intense winter cooling and excessive summer heating. Annual temperature ranges of 50°C and absolute temperature ranges 80—90° are recorded in Mongolia. In the temperate latitudes Grigorev's and Budyko's (1956) radiational index of dryness increases rapidly from below 1 to above 3. Consequently, the semi-desert and steppe zones expanded north of 50°N, although both the inland basins and plateaus occur at high altitudes (1000—1800 m asl.). At the same time the permafrost shield stretched southward to about 46°N (Gravis 1974). Because of water shortage the "dry" permafrost type is dominant leading to the granular distintegration of bedrock and to deflation. As a result of these "zonal" migrations, the catenas present a variety of patterns which are due to the availability of moisture. Kowalkowski and Pacyna (1977) have illustrated this for the forest-steppe belt where the dry, southern slopes are clearly contrasted with the moist, northern slopes carrying patches of forest associations fed by melting permafrost.

A fourth control producing the geoecological belts are the high-mountain barriers which enclose the vast plateau of interior Asia. Mountains effectively stop the advected moisture from penetrating inland, and in these areas the largest temperature inversions in the world can be observed (Voyeykov 1895). The Pamir, the Himalayas, the Tibetan Plateau and the mountain ranges of East China completely block the summer monsoon. High precipitation and thermal gradients are to be found on the western border of the Pacific (Nikolskaya 1974).

On the northern fringe of the extensive semi-desert inland basins of Mongolia, the Khangai and the Mongolian Altai Mts receive more precipitation and their climate is less continental. The existing zones are different from those in the Kun-lun' Mts, and even in the isolated Gobi Altai mountain blocks. According to Yunatov (1950), this is the sub-humid type of altitudinal zonation due to a milder type of climate for under long-continued inversion conditions in the winter a warm layer exists at higher levels in the mountains (Chelpanova 1963).

Within the wide inland basins the dry air masses which become extremely hot in summer and cold in winter tend to affect the surrounding mountains. For this reason, the southernmost parts of the mountains experience a drier climate, and the variety of habitats is reduced in number there (Karamysheva and Banzragch 1977).

In the mountains of Central Asia the four factors discussed above, i.e. the zonally arranged radiation and air pressure balances, the distribution of land and of mountain ranges produce various types of geographical belts which are unique on a worldwide scale. Paleogeographical evidence including erosion surfaces (Ravsky 1972), and the evolution of both flora and fauna seems to suggest that the above types of altitudinal zones remained relatively constant over the last million years. There was, however, a shift in the belts during the Pleistocene. Under cool-continental conditions the principal cause of environmental changes affecting both the mountains and the basins of interior Asia were the moisture-controlled variations, since the geoecological systems show a consistent water deficit. An unimportant disturbance of the water balance would cause shifts in both zones and belts.

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ЗОНАЛЬНОСТЬ ЮЖНОГО СКЛОНА ХАНГАЯ (МОНГОЛИЯ)

Резюме

К. Климек

ГЛАВНЫЕ ФИЗИКО-ГЕОГРАФИЧЕСКИЕ ЧЕРТЫ ЮЖНОГО СКЛОНА ХАНГАЯ

Горы Хангай расположены на границе двух крупных структурно-орографических единиц: гор Южной и Восточной Сибири, а также равнин Центральной Азии. Главный хребет Хангая, длиной ок. 700 км, тянется с NW на SE. Он возвышается в среднем до 3000—3500 м н. у. м., а самая высокая вершина Отгон-тенгри достигает 4031 м н. у. м. Южные предгорья Хангая представляют собой холмисто-волнистую Южнохангайскую возвышенность. Ее поверхность понижается к югу от 2700 м н. у. м. у подножья Хангая и до 1500 м н. у. м. на границе Гоби.

Южный склон Хангая в структурно-палеографическом и геоморфологическом отношениях ярко выделяется как особая тектоническая единица. В ней наблюдается чередование горных южносибирских ландшафтов и плоскогорий Гоби.

Такие элементы природной среды как климат, воды, а также растительный покров изменяются очень быстро по мере изменения высоты. В результате этого в поле зрения находятся одновременно типичные ландшафты высокогорной тундры с пятнами вечного снега и полупустынные ландшафты.

Из-за большой „массивности" гор и их большой относительной высоты их южный склон представляет собой обособленный пояс, разделяющий Южную Сибирь от Центральной Азии в климатическом, гидрологическом, почвенном и геоботаническом отношениях. У его подножий проходит глубокий разлом земной коры, разделяющий две структурно-тектонические единицы, обладающие разным геологическим строением и различно развивавшиеся в геологическом прошлом. Этот тектонический пояс был еще активным в плейстоцене. Существуют доказательства, что тектонические движения продолжаются и в настоящее время.

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ИСТОРИЯ ФИЗИКО-ГЕОГРАФИЧЕСКИХ ИССЛЕДОВАНИЙ В ГОРАХ ХАНГАЙ

В горах Хангай с незапамятных времен обитали кочевники. Доступность горных долин, обилие воды и кормов, несмотря на суровые климатические условия, благоприятствовали животноводству. Здесь жили племена гунов, тюрков, а потом монголов. Легкопередвигающиеся кочевники с давних времен хорошо знали природу этих гор, они используют эти знания и в настоящее время.

Первые информации о природе этой территории начали проникать к индоевропейским народам только в XIII веке, но лишь в XIX веке начался период систематических исследований природы Монголии. Они охватили также горы Хангай. Особые заслуги в изучении района принадлежат систематическим экспедициям русско-географического общества.

Образование Монгольской Народной Республики (1921) создало возможность развития монгольских научных учреждений, среди них Академии наук МНР. Основание Института географии и мерзлотоведения АН МНР, а также необходимость изучения природных ресурсов для потребностей народного хозяйства привели к развитию физико-географических исследований в Монголии. Кроме монгольских специалистов в исследованиях принимают участие также ученые других стран, главным образом Советского Союза.

Совместная монголо-польская физико-географическая экспедиция, организованная Институтом географии Польской академии наук и Институтом географии и мерзлотоведения Монгольской академии наук, вела исследования на южном склоне Хангая в 1974—1975 годах. Целью исследований было комплексное изучение физико-географических условий бассейна Ульдзеиту-гол. До настоящего времени было опубликовано несколько работ, которые заключают в себе первые оригинальные наблюдения, касающиеся физико-географических проблем этого района.

Настоящий том заключает ряд статей, в которых авторы пытаются высказать свою точку зрения на проблемы природной среды южного склона Хангая в аспекте её вертикальной зональности.

К. Климек

РЕЛЬЕФ И ПАЛЕОГЕОГРАФИЯ ЮЖНОГО СКЛОНА ХАНГАЯ

В рельефе южного склона Хангая были выделены внутренняя и внешняя зоны. Во внутренней зоне, шириной в 10—20 км, преобладают обширные плоские поверхности с врезанными в них глубокими цирками или ледниковыми долинами. Внешняя зона, шириной в 20 км, расчленена глубокими узкими речными долинами на несколько изолированных горных массивов.

На этой территории выделены верховые поверхности выравнивания (3400 м н. у. м.) и надбазисные поверхности выравнивания (3100—3300 м н. у. м.), а также базисные поверхности выравнивания (2650—2850 м н. у. м.). Нижнемеловые коррелятивные отложения, выступающие на восточном окаймлении Хангая, как и верхнетретичные базальтовые лавы, покрывающие верховые поверхности выравнивания, позволяют предполагать, что только активизация тектонических движений третичного периода дала начало формированию этого горного массива. Значительно моложе тектонические впадины у южных подножий Хангая. Их образование вызвало большие изменения базиса эрозии, что явилось импульсом усиления последней и интенсивного расчленения внешней зоны гор.

Последнее оледенение этой территории имело долинный, двуфазовый характер. В конце последнего оледенения климат отличался большой сухостью. Недостаток атмосферных осадков вызвал отмирание долинных ледников. Водно-ледниковые отложения переходят у подножья гор в озёрные осадки, заполняющие тектонические котловины. Возраст органических остатков, найденных в кровле этих отложений, определен в 17 220±155 лет B. P.

Е. Бжезняк, Т. Недзведзь

ВЕРТИКАЛЬНАЯ ЗОНАЛЬНОСТЬ КЛИМАТА

На основании данных многолетних наблюдений и результатов стационарных измерений, проведенных во время физико-географических экспедиций в летний период 1974 и 1975 годов, представлена пространственная дифференциация основных элементов и показателей климата на южном склоне Хангая.

В этом районе наблюдается типично горно-долинная циркуляция; направления течений воздушных потоков обусловлены системой форм рельефа. Особое внимание было обращено на различия температуры на разных уровнях, а также на интенсивность инверсии температуры воздуха. Определялись также изменения в зависимости от высоты, температуры поверхности грунта и почвы, влажности воздуха, облачности и атмосферных осадков.

Тесная зависимость температуры июля от высоты над уровнем моря дала возможность, используя метод классификации М. Гесса (1965), различать на южном склоне Хангая шесть термических вертикальных зон: степей, нижней лесостепи, верхней лесостепи, нижних высокогорных лугов, верхних высокогорных лугов и высокогорных каменистых лугов. Установлено, что верхняя граница леса на территории южного Хангая находится на уровне верхней границы холодной зоны (2700 м н. у. м.) и связана с изотермой июля +9°C.

Р. Соя

ПОВЕРХНОСТНЫЕ ВОДЫ ЮЖНОГО СКЛОНА ХАНГАЯ В БАССЕЙНЕ ЦАГАН-ТУРУТУИН-ГОЛ

Дифференциация вод в бассейне реки Цаган-Турутуни-гол носит черты зональности. Изменяющаяся вместе с высотой над уровнем моря величина стока из бассейна, изменения других гидрологических параметров являются отражением климатической зональности этой территории и зональной дифференциации многолетней мерзлоты. В бассейне реки Цаган-Турутуни-гол можно выделить две территории: предгорье Хангая с отрицательным водным балансом и горную территорию с положительным водным балансом. Границей между двумя территориями является морфологическая граница гор.

В горном районе, где существуют условия для формирования постоянного стока, его величина в летнее время обусловлена средней высотой бассейна над уровнем моря. Бассейны со средней высотой выше 2800 м н. у. м., занимающие около 32% территории изучаемого района с положительным водным балансом, дали во время исследований около 60% полного стока. Бассейны со средней высотой 2500—2800 м н. у. м., занимающие около 45% территории, дали около 30% стока. Бассейны со средней высотой ниже 2500 м н. у. м., занимающие 23% территории, дали только 10% стока. В части бассейна реки Цаган-Турутуин-гол, с отрицательным водным балансом, происходило уменьшение стока, достигающее почти 20%. На территории с отрицательным водным балансом поверхностные воды (маленькие озера, заболоченности) связаны с речными водами или водами тающей многолетней мерзлоты. На территории с положительным водным балансом существование маленьких озёр и заболоченностей зависит от атмосферных осадков и в большой степени от рельефа местности.

К. Пенкаля, Т. Зентара

СОВРЕМЕННОЕ ФОРМИРОВАНИЕ СКЛОНОВ В ЮЖНОМ ХАНГАЕ

На южном склоне Хангая заметно выделяются морфогенетические зоны:

— тундры и высокогорных лугов, расположенных выше 2700 м н. у. м. На склонах с северной экспозицией высокогорная тундра спускается ниже, до высоты 2700 м н. у. м.;

— лесостепи (горного леса и степи) от 2100 м н. у. м. до 2700 м н. у. м.;

— степи до высоты 2700 м н. у. м.

У подножья гор рельеф степных районов формируется в настоящее время под влиянием смыва, эпизодической эрозии, суффозии и дефляции. Местами на низких, более влажных террасах развиваются процессы морозной сегрегации, пучения и образования морозобойных трещин.

В зоне лесостепи самую большую роль играют поверхностный и линейный смыв, возрастает роль эпизодической эрозии и дефляции главным образом на возвышенностях. Периодами, при участие многолетней мерзлоты, происходит сползание. Менее интенсивны такие процессы как: солифлюкция, суффозия и корразия.

В зоне тундры и высокогорных лугов большую динамику проявляют процессы, связанные с глубиной залегания многолетнего мерзлотного слоя. Решающую роль в формировании возвышенностей, склонов и дна долин играют физическое выветривание, процессы морозной сегрегации, пучение, напряжения при образовании морозобойных трещин, процессы солифлюкции, смыва и гравитационных склоновых движений, а также процессы нивации, дефляции и корразии.

Соотношение этих процессов, а также их большая динамика ведут к образованию криопланационных террас и поверхностей, а также ступенчатого профиля склонов. Как на возвышенностях, так и на дне долин развиваются структурные грунты, находящиеся в разных стадиях развития.

Перигляциальные склоны, в первом этапе развития, являются неровными — их поверхности в результате процессов, действующих согласно их уклону, т. е. вниз и вверх, а также процессов бокового срезания неровностей зависят от геологического строения (граниты, базальты).

С точки зрения зональности и динамики современных криогенных процессов криопланационные террасы и горизонты южного склона Хангая можно разделить на очень активные, активные малоактивные, фоссильные и субфоссильные. Криопланационные террасы и горизонты в пределах Центрального Хангая располагаются выше 3200 м н. у. м. Активные террасы на склонах с южной экспозиции находятся выше 2900 м, а на северных выше 2800 м. Малоактивные террасы находятся на высоте выше 2700 м и представляют собой переходный этап к фоссильным формам. Фоссильные террасы располагаются в зоне высокогорных лугов от 2400—2700 м н. у. м.

Большую роль в развитии склонов в Хангае играет экспозиция, от которой зависит мощность активного слоя многолетней мерзлоты, сочетание современных процессов, а также их ритм и интенсивность. В настоящее время наблюдается их влияние на различное формирование склонов.

А. Ковальковски

ВЕРТИКАЛЬНАЯ ЗОНАЛЬНОСТЬ ПОЧВ ЮЖНОГО СКЛОНА ХАНГАЯ

На южном и западном склонах Хангая наблюдались разные варианты почвенных зон. Их возникновение связано с высотой над уровнем моря, а также с перемещением с юга на запад в глубь горного массива масс воздуха из Долины великих озёр. На изучаемой территории гидротермические условия криогумидной среды на севере, переходящей в криоаридную на юге, являются, несомненно, главными факторами, определяющими асимметрию высотной зональности почв, а также инверсию зон криогумидно-аридного типа.

Ярко выражены границы зон в криогумидной части выше 2700 м н. у. м. Быстро возрастающая к югу криоаридность является одним из факторов затушевывания высотной зональности почв. Каждая высотная зона характеризуется определёнными чертами взаимодействия морфогенетических и почвообразовательных процессов, придающих ей черты географического своеобразия (табл. 1, 2, 3). Фактор высоты над уровнем моря является безусловно изменяющимся во времени и в условии смены климатов. Сталкивающиеся и передвигающиеся действия криогумидности с доминирующими процессами криогенного бурого выветривания с севера и криоаридности с доминирующим каштановым процессом с юга, вызывает то, что рядом образуются контуры полигенетических и реликтовых почв определённых хроносеквенций педогенезиса.

Ретардативные свойства почв несомненно влияют на экологические условия роста растений. Растительные же сообщества легче приспосабливаются к изменённым климатическим условиям. Потому нельзя их считать одним из комплекса показателей вертикальной зональности почв.

А. Пацина

РАСТИТЕЛЬНОСТЬ ЮЖНОГО СКЛОНА ХАНГАЯ И ЕГО ЗОНЫ

Хангай является переходной зоной между таёжной Восточной Сибирью и степной Центральной Азией. На южном склоне Центрального Хангая образовалась аридная система растительных зон (южнохангайская; Карамышева, Банзрагх 1976) со следующими зонами: степи, лесостепи и высокогорной зоной, в которой можно различить две подзоны — нижнюю и верхнюю.

Зона степей занимает небольшие пространства и представляет собой ковыльно-полынную степь (с *Stipa Krylovii* Roshev, *Artemisia frigida* Willd. ; таб. 8, сн. 5, 6).

Лесостепная зона (*sensu* Юнатов 1950) занимает большую часть территории. Лиственничные леса (псевдотаёжного типа; Коротков 1976) занимают только небольшой процент территории (таб. 9). Они занимают отдельные пространства на склонах с С, ССВ и ССЗ экспозициями. Их флористический состав показывает таблица 9. Остальную территорию зоны лесостепи занимают горные степи, изменяющиеся в зависимости от условий местообитания (таб. 8, сн. 1—4, 7—10).

Аридный климат сильно повлиял на образование высокогорной зоны. В ней отсутствуют псдтольцовое редколесье и горная тундра кустарного, травянистого, мохового или мохово--лишайникового типов, характерные для гор Южной и Восточной Сибири (Карамышева, Банзрагх 1977). Горные склоны в нижней высокогорной подзоне занимают кобрезовые луга, свойственные горам Центральной Азии (Карамышева, Банзрагх 1976а, 1977), а дно — заболоченные высокогорные луга. Верхняя высокогорная подзона сильно контрастирует с более низкими горными зонами и степями южного подножия Хангая. Склоны заняты каменными россыпями, а плоские, широкие хребты и высокие плато — каменистой горной тундрой, которая по условиям местообитания и по морфологии близка типу арктической, каменистой горной тундры. Растительностью покрыт только небольшой процент территории. Среди скудной флоры сосудистых растений (ок. 25 видов), можно выделить высокогорные сибирские и высокогорные арктические виды.

Л. Старкель

ЗОНАЛЬНОСТЬ В ГОРАХ С КОНТИНЕНТАЛЬНЫМ КЛИМАТОМ

Причиной образования геоэкологических зон гор в континентальном климате являются экстремальные величины режима термики и влажности, которые определяют ход физических, химических и биологических процессов. Внутри Азии на расположение этих зон, на проникновение полупустынных районов далеко к северу, а одновременно проникновение далеко к югу вечной мерзлоты и зоны лесов повлияли четыре фактора: радиационный баланс и рост температуры по направлению к экватору, уменьшение атмосферных осадков по направлению к тропической зоне, величина континента и удаление его центра от океана, а также явление инверсии в обширных понижениях среди гор. Окаймления сухих понижений, к которым принадлежит южный склон Хангая проявляют значительный градиент влажности, обусловливающий чередование зон на небольшой территории.

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VERTICAL ZONALITY IN THE SOUTHERN KHANGAI M UNTAINS