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LESZEK STARKEL (*)

SHIFTING OF CLIMATIC-VEGETATION BELTS IN EURASIAN MOUNTAINS AND THEIR EXPRESSION IN SLOPE EVOLUTION

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Climatic, vegetation vertical zones controlled mainly by changes in the temperature, shifting in the Quaternary, decide on the rate and direction of slope evolution. In the Eurasian Mountains beside vertical displacement of geomorphic processes during cold stages a pronounced role was played by expansion of permafrost in western direction. The limits of vertical zones of geomorphic processes are frequently adapted to inherited features of relief and especially to relief energy. High rate of uplift cause accelerated incision of valleys and large slope failures, it may be more important than the secular processes acting in various climatic vertical zones. The mountain glaciers during advances separate valley sides into supraglacial, paraglacial and subglacial sectors.

KEY WORDS: Morphoclimatic vertical zones (belts), Slope evolution, Glacial-interglacial shift of Vertical zones, Relief energy, Uplift and downcutting, Transforming role of mountain glaciers.

INTRODUCTION

High mountains are the product of tectonic uplift and consequent exposure to degradation. The mountain slope, the main element of mountain relief, is the arena of transfer of water and mineral matter. From the ridge crest to the valley bottom there is an associated increase in energy and sediment load. Slope evolution is controlled by two climatic factors, temperature and precipitation. In their vertical distribution the boundary role is played by passing of freezing point (0°C).

Beside the role played by these forces in determining the variety in present types of slope evolution, a substantial role is played by existing forms inherited from the past developed on various geological substrates. Among these features are shape, gradient and height of slope, rock resistance and especially the glacial and crionival features created during the Pleistocene. Therefore the real shapes of mountain slopes and mosaic pattern of their elements may be far from the ideal slope profile presented by Dalrymple & *alii* (1968).

But in the evolution of high mountains the character and stability of slope base level is also important. Base level can be represented by a wide terraced valley floor, a changing surface of valley glaciers with paraglacial deposits or, in tectonically active ranges, an entrenched river channel, transporting downstream a bedload delivered from slopes by rockfalls or debris flows.

The relief energy determines the contrast between ridge crest and valley floor, which may pass 1000-2000 m and is reflected in lowering of temperature with rising altitude and in the development of several climatic-vegetation belts. Change in temperature and precipitation is in reality continuous, even Höllermann (1973) speaks of «abstract altitudinal belts». But there are the boundary conditions (or factors) like heat balance of the ground, number of frost days, duration of vegetation season, precipitation-evaporation relationship etc. which limit the existence of various processes especially biochemical ones influencing vertical soil types and the extent of plant species and communities.

All these factors create a platform for distinguishing morphoclimatic vertical belts. Their number and sequence depend both on the relief energy and location within one of the climatic zones.

Frequently in our studies we forget about vertical zonation, taking as one unit the whole high mountain relief rising above present-day timberline or the whole landscape left after extensive Pleistocene glaciation (cf. Troll, 1973b).

(*) *Department of Geomorphology and Hydrology, Institute of Geography, Polish Academy of Sciences - 31-018 Kraków, ul. św.Jana 22.*

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The knowledge of this zonality in type and rate of geomorphic processes is rather fragmentary, being restricted either to part of slope profile like talus or alluvial fans (higher rocky segments are considered as a source of deposits) or to selected processes like solifluction, debris flow, patterned grounds, and avalanches. In the second case, their altitudinal extent helps to separate morphodynamic vertical zones. But the spatial extent of phenomena such as solifluction lobes or patterned grounds depends on the altitudinal position of flat or steeper slope fragments over which such processes occur.

Very rare are the studies considering the evolution of the whole slope profile in relation to morphoclimatic vertical zonality (Caine, 1982; Kotarba, 1984; 1988) or even the calculation of total denudation balance in small basins located in one vertical zone (Rapp, 1960).

When discussing the role of shifting of climatic vertical zones in the mountains we should not forget that a leading role in the high mountains may be played by the tectonic factor (Shroder & Bishop, 1998) or in mature landscapes of foothills only processes connected with zonal climatic parameters may operate (Starkel, 1969).

MORPHOCLIMATIC VERTICAL ZONES

The dominant geomorphic processes in mountains are controlled mainly by temperature. The mean annual temperature goes down about 0,5-0,8°C for every 100 m elevation. The most important temperature parameters are the annual isotherm +2°C and July temperature +10°C, which coincide with upper timberline in the European mountains (fig. 1). There are several thresholds, determined by the circulation of water, physical and chemical weathering and finally on the character of vegetation, soil and transformation of slopes. Among them are the orographic snowline, the lower limit of permafrost, lower extent of solifluction and other cryogenic processes, upper timberline, lower timberline and in arid zones, the steppe-desert ecotone (Troll, 1973 b; Wilhelmy, 1974).

The snowline in European mountains coincides with mean annual temperature –2°C and separates the glacial belt favourable for the formation of ice fields and lower cryonival belt (frequently called periglacial), characterised by seasonal melting of snow and cryogenic processes. The latter may go to the upper timberline located 600-1000 m below snowline. In cryonival belt are two distinct parts: the upper one with blockfields and patterned grounds bordered by permafrost limit and the lower one with solifluction lobes and small size structural soils covered by subalpine vegetation (Furrer, 1975; Karrasch, 1977; Matsuoka, 2003; Kotarba, 1980; Rączkowska, 2007 - fig. 1).

The permafrost limit coincides usually with mean annual temperature about 0 to –1°C but the altitudinal extent below snowline is very variable depending on aridity, reaching even 1400 m (Matsuoka, 2003).

The upper timberline has a character of a wide transitional zone. In the European mountains, forest is climbing up in the elevated convex slope parts and lowering on free-face wall and in slope depressions affected by avalanches (Troll, 1973a). A totally opposite behaviour of the timberline is observed in continental climate of Central Asia, where in moist slope depressions the trees are shifting higher (Starkel, 1994).

The wide forest belt in area of humid climate is characterised by surplus of water and biochemical processes as well as by mass movements, piping and linear erosion (depending on slope gradient and permeability of substrate - cf. Starkel, 1959). The forest belt is blocking the transfer of load carried from upper crionival zone.

The lower timberline is controlled in turn by surplus of evaporation over precipitation. Below that line the role of eolian denudation and rare heavy downpours becomes more important.

The limits between vertical zones are crossed during the extreme events like heavy rainfalls, snowfalls or rapid temperature rises. During heavy rains the debris flows created above the forest belt enter into the forest, as happened in Tatra Mts in July 1997 (Kotarba, 1998). The effect of warming and heavy rain combined with snowmelt in

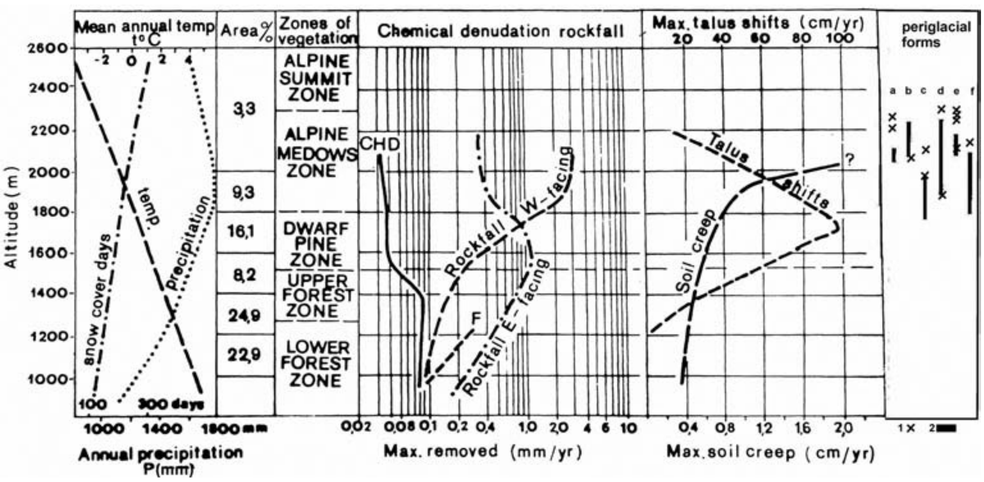

FIG. 1 - Differentiation of some geomorphic processes in vertical climatic zones of the Tatra Mts (after Kotarba 1984 and Rączkowska 2007). Periglacial forms: a. sorted circles, b. polygons, c. gelifluction lobes, d. nival niches, e. protalus ramparts, f. rock glaciers. 1. altitudinal extent, 2. active forms.

French Alps has been described by Tricart & *alii*, (1962). In Central Asia, the rapid advections of humid air from the Atlantic cause rapid snowmelts reaching to 4500 m asl. combined with debris flows and floods (Starkel, 1976). Such events have been described in Tianshan and Pamir by Gontscharov (1962) and many others. Conversely, heavy snowfalls can cause large avalanches in the forest belt.

TYPES OF VERTICAL ZONALITY

In the Eurasion mountains at least 3-4 different sequences of vertical zonality may be distinguished, illustrated on the latitudinal and longitudinal transects of Eurasia (figg. 2-4).

In the oceanic climate of west-European mountains, both upper timberline and snowline descend in the west coastal zones exposed to higher precipitation and strong winds. In Scandinavia the snowline descends to 1000 m asl. And the treeline approaches sea level in the north (Sondersen, 1997). In Britain, after deforestation the former woodlands have been replaced by blanket bogs, because it was to wet for tree growth (Moore, 1983).

In areas with a continental climate, the position of snowline and timberline is much higher due to deficit of precipitation. Therefore at lower elevations the steppe zone and even the desert are present (fig. 2). In central and northern Asian highlands, the mean annual temperature drops several degrees below 0°C., causing the formation of permafrost and storage of water in snow and ice (Serebryanny & Gravis, 1993; Starkel, 1998 - fig. 4). The radiation aridity index in the range 1-3 extends along most the Asian transect (Budyko, 1971). Therefore steppe and deserts extend from subtropics up to the boreal zone. In Mongolia the forest belt occupies only narrow belt (300-500 m) in the middle part of slopes limited not only by temperature (upper limit) and rainfall (lower limit) but also by discontinuous permafrost. Its active layer is feeding the coniferous forest with water (Haase & *alii*, 1964; Kowalkowski & Starkel, 1984). This is well exemplified by sequence of soils in Khangai Mts. (fig. 5), where the forest and chernozem belt (2300-2600 m asl.) separate the lower cryosemiarid zone extending south to the Gobi desert and the upper cryosemihumid zone sloping down towards the East-Siberian mountains (Kowalkowski & Starkel, 1984; Shahgodaneva, 2002). In that cryosemihumid zone from 2800 m asl. to the snowline extends the area of cryoplanation terraces (Pękala & Repelowska-Pękala, 1993). Towards south on the Tibetan Plateau elevated to 5000 m asl. (with mountain ranges rising about 2000 m above it) the permafrost occupies areas above 4000-5000 m and a snow-

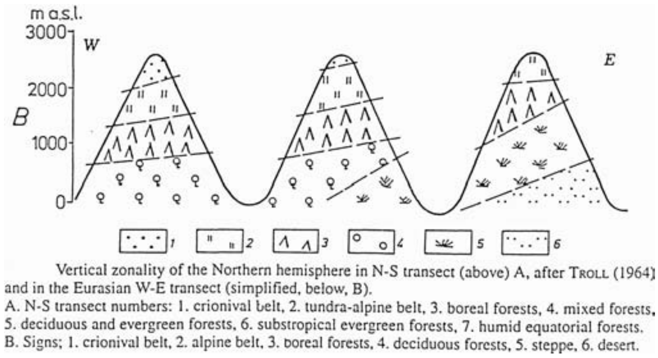

FIG. 2 - W-E Eurasian transect of vertical zonality in temperate zone (after Starkel 1993). 1. cryonival belt, 2. alpine belt, 3. boreal forest, 4. deciduous forest, 5. steppe, 6. desert.

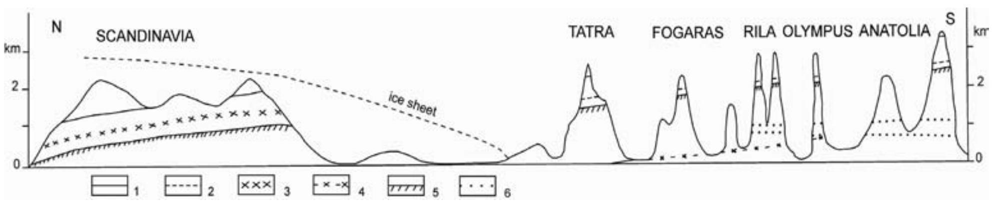

FIG. 3 - N-S European transect of vertical zonality during last cold stage and at present. 1. present snowline, 2. cold stage snowline, 3. present lower limit of permafrost, 4. cold stage limit of permafrost, 5. present upper timberline, 6. last cold stage upper and lower timberline.

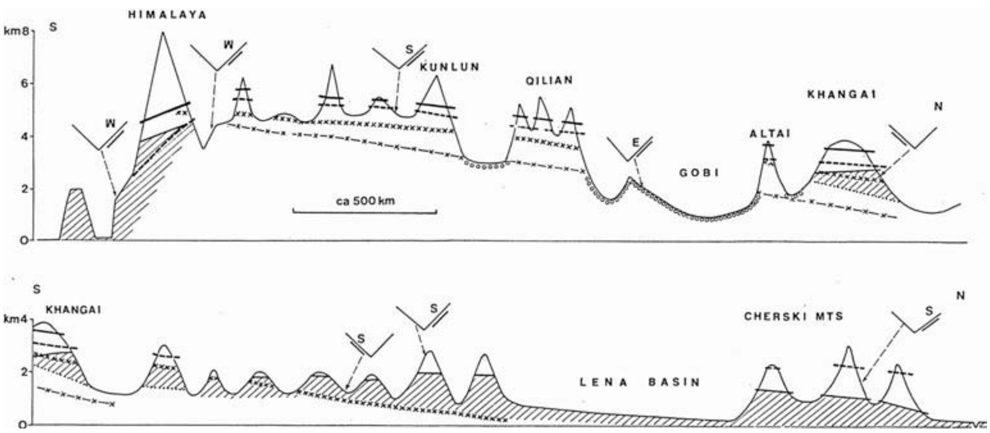

FIG. 4 - S-N Asian transect of vertical zonality during last cold stage and at present (after Starkel 1998). 1. present snowline, 2. cold stage snowline, 3. present lower limit of permafrost, 4. cold stage limit of permafrost, 5. present upper timberline, 6. present lower timberline, 7. asymmetry of slope processes (more dominant processes: S - glacifluction, W - slope wash, E - eolian), 8. present desert.

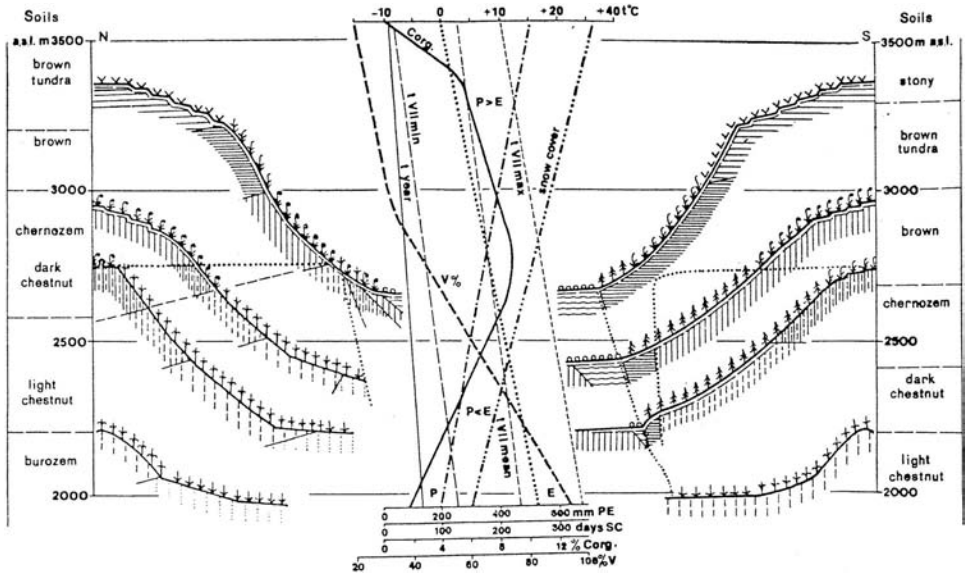

FIG. 5 - Vertical zonality on valley sides in S-Khangai Mts (Kowalkowski and Starkel 1984), t - temperature, P - precipitation, E - evaporation, C org. - carbon content, 1-5 cryosemihumid vertical zone (1. rocky tundra initial brown soils, 2. Kobresia high mountain meadow brown and grey-brown soils. 3. brownsoils with larch forest-steppe, 4. Kobresia high mountain meadow steppe, brown and hydrogenic chernozems, 5. relic brown chernozems with larch forest-steppe), 6-11 cryosemiarid vertical zone (6. chestnuit chernozems with forest steppe. 7dark chestnut soils with forest steppe, 8. chestnut chernozems with dry steppe, 9. dark chestnut soils with dry steppe, 10. light chestnut soils with dry steppe, 11. burozems of dry steppe), 12. cryohumid swampy soils of mountain meadows (intrazonal soils), 13. permafrost, 14. limits between different zones and asymmetry of habitats on slopes and in valley bottoms.

line is only 1 km higher. The southern slope of Himalayan range and SE margin of Tibetan Plateau are invaded by summer monsoon rains, the present snowline descends rapidly from about 5800 to 4500 m asl. and forests in Eastern Himalaya climb to 3500-3800 m asl. (Kalvoda, 1984 - fig. 4). Precipitation at the mountain edge at lower altitudes reach 4000-6000 mm.

MESOCLIMATIC VERTICAL ZONES

The mountain slopes in the narrow valleys and at the margin of intramontanous basins show additional mesoclimatic differenciation reflected also in geomorphic processes. There are three vertical mesoclimatic belts (cf. Yoshino, 1984; Obrębska-Starkel, 1970). The lower sector of valley floor rising to 100-150 m above rivers is characterised by inversion of temperature, high humidity and frequent night frost. This is reflected in intense frost weathering and formation of talus fans below rock-cliffs, known from karstic canyons (Ložek, 1975). The inversion in E-Siberian intramontanous depressions is so high that their bottoms are covered by extensive icings (Starkel, 1998). The middle parts of slopes show lower amplitudes of temperature, this warm belt has preserved relict plant species from warmer periods. The ridges are cooler again, being exposed for advection of fresh air masses and strong winds, causing deflation and lowering of upper tree line (Holtmeier, 1993).

SLOPE EXPOSURE AND ASYMMETRY

The various aspects of entire mountain ranges as well as of particular slopes in mountain valleys determine the differences in insolation, temperature and also in precipitation and humidity. Therefore the type and intensity of processes active on opposite slopes in one vertical zone may differ greatly.

In the scale of mountain ranges we observe a great contrast in position of permafrost limit and snowline between N-facing and S-facing slopes in the Altai, Sayan or Khangai Mts. of Central Asia (Schukin, 1950; Shahgodaneva, 2002) as well in the whole Tibetan Plateau with surrounding mountains (fig. 4 - Hövermann & Lehmkuhl, 1993; Starkel, 1998). But much more distinct is the aspect exposure towards the humid air masses in the temperate zone from the west and in the monsoonal Asia from south (fig. 4), being especially reflected in the position of tree line. In the Pyrenees, the differences between humid NW side and drier S-side reach 500-700 m. The precipitation at the margin of Alps is much higher than in the intramontane basins (Troll, 1973a) being reflected in the elevation of upper timberline, fluctuating between 1600 m asl. on the north, 2000 m on the south and upto 2300 m in the central part. In Caucasus, the timberline rises from 2200 m in the west to 2600 m in the east (Schukin, 1950). Great contrast exists between two sides of Hindukush (Ratjens, 1972) and on the S-margin of Himalaya where at distance of about 50 km annual rainfall decline from 4000-6000 mm to below 1000 mm (Bailie & Norbu, 2004). At the southern scarp of the Meghalaya Plateau the contrast is even greater and the highest rainfall of more than 11000 mm is recorded in the middle of scarp at the level of condensation during process of convection (Starkel & Singh eds., 2004).

In the scale of particular valley slopes the most distinct difference is between opposite N and S-facing slopes, which are characterised by differences in insolation and temperature. In oceanic climate of temperate zone in the forest belts, the differences are not so great and usually do not exceed 200-300 m in case of upper tree line (Troll, 1973a). In arid climates, such differences are controlled by the water balance. For S-facing slopes the lower forest limit is located higher by 500 m and more. In central Hindukush all vertical belts are located higher on S-exposed slope, being evident in geomorphic processes (Kaszowski, 1984). In Khangai Mts. on drier slopes, the steppe zone is rising 400 m higher and passes directly to alpine belt (Kowalkowski & Starkel, 1984 - fig. 5). Only on N-facing slopes with discontinuous permafrost at elevation 2200-2600 m asl. are larch forests growing, fed by water from active layer, which also facilitates gelifluction. In upper parts of slopes, cryoplanation is a common process. In the Siberian mountains of the boreal zone under continuous permafrost, the slopes show an opposite asymmetry. The S-facing slopes both in forest belt and above it are modelled by gelifluction due to a longer melting season (Baulin & *alii*, 1984).

INHERITED ELEMENTS OF MOUNTAIN SLOPES

The climate-vegetation vertical zones are superimposed on inherited landscape. The origin and history of relief as well its geological structure (lithology and tectonics of bedrock) determines the shape and inclination of slopes as well as the depth of river valleys and the contrast in vertical zonality of particular slopes.

Among main types of inherited relief we may distinguish dissected mature or planated relief, young V-shaped mountain relief, mature structure controlled relief, landscapes transformed by mountain glaciers and ice sheets or by long-lasting periglacial processes (fig. 6).

The uplifted and deeply dissected mature relief is characteristic of old rejuvenated horst mountains, typical for northern Europe and central Asia. The Scandinavian fjelds, highlands of Scotland and many central Asian ranges were planated before Quaternary and later lifted up to glacial or cryonival alpine belts (cf. Wilhelmy, 1974; Rapp, 1960; Rączkowska, 2007).

The young mountain landscapes mainly typical for the alpine orogenic system are characterised by deep V-shaped valleys with steep slopes and knife-edge ridges and in tectonically very active areas modelled by slope failures (Shroder, 1998; Fort, 2003) exposing cliffs built of resistant rocks. The entrenching of deep Quaternary valleys followed in steps formed during cold stages of prevailing aggradation (Starkel, 1969).

The structure-controlled relief is typical for folded mountains built of rocks of different resistance like flysch belt characteristic for Alps and Carpathians (Starkel, 1969). Usually more resistant sandstone or limestone beds form residual ridges with steep slopes which at present or during cold stages modified the shifting of vegetation vertical zones. (Panizza & *alii*, 1997). Their limits coincide with change in rock resistance and circulation of water. In high mountains many changes in slope shape have been modelled by valley glaciers and large ice sheets. The relief of cirques surrounded by knife-edge nunataks, glacial troughs and their continuation in fjords deliver in mountain valleys a great number of vertical rocky walls which at present are located in alpine or even in forest belt (cf. Troll, 1973a; Kotarba, 1988). Similar transformations occurred in low mountain landscapes affected by periglacial processes. The flat surfaces of cryoplanation origin may be found as terraced watersheds as well at the foot of slopes as cryopediments (cf. Kowalkowski & Starkel, 1984).

These older fragments of various origin are incorporated in the climatically controlled vertical zonality on mountain slopes and may form cliffs, steep or gentle, rocky or debris segments in different altitudinal belts of mountain landscape.

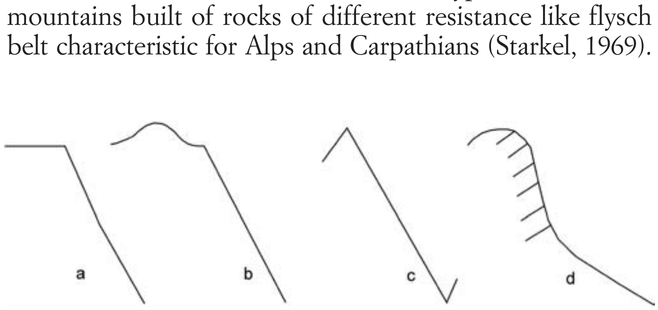

FIG. 6 - Different origin and history of mountains slopes: a/ planation surface dissected by deep valleys, b/ rejuvenated mature landscape, c/ young mountain relief, d/ mature structure controlled slope.

VERTICAL ZONALITY IN GLACIATED AREAS

The mountain glaciers are the product of glacial zone above the snowline. These may have a character of fjeld ice fields over planated highlands (which may reach a scale of ice sheets) or valley glaciers flowing down valley. The glacial erosion transforms the preglacial relief causing overdeeping of mature valley heads or lateral extension of V-shaped fluvial valleys leading to formation of glacial troughs (fig. 7). In high mountains with relief energy to 2000-3000 m, the main valley glacier is supplied frequently by hanging tributary glaciers or by avalanches. The overdeeping by great ice masses flowing from extensive ice sheets (like in Scandinavian mountains) may reach hundreds meters and form fjords with vertical cliffs.

The mountain glaciers divide the former uniform slopes into separate parts. Upper slope sectors represent a supraglacial zone rising above ice surface with nunataks and are located in the cryonival belt or above the snowline (Kellerer-Pirklbauer, 2008). This zone may even reach forest belt, if the glacier goes so far downstream. On the

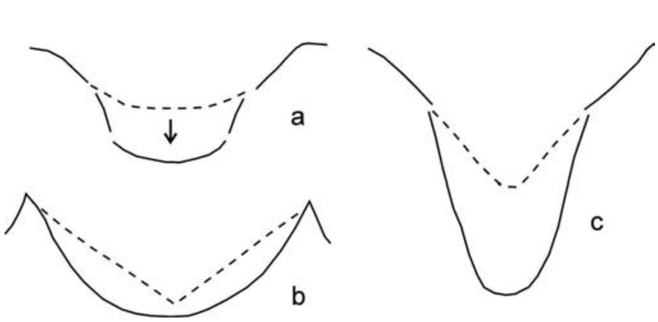

FIG. 7 - Glacial erosion transforming older river valleys: a/ wide mature river valley, b/ V-shape young river valley, c/ large mountain valley transformed in deep fjord valley.

flanks of glacial tongues is a paraglacial belt of aggradation which consist of debris covering ice surface, talus, lateral moraines, nivation ramparts and the rockglaciers in case of retreating glaciers (Kääb, 2007; Luckman, 2007). Below that level there can be a deep trough filled during advance by ice. Frequently that paraglacial zone from cold stage can be still the base level of slope processes after deglaciation (Kotarba, 2008 - fig. 8).

At the front of glacier extends the proglacial zone, which is expanding up-or down valley in relation to advancing or retreating front of glacier. The mountain glaciers during cold stages of Pleistocene moved 1000-2000 m and more below ice fields and now the lower segments of slopes exposed to denudation are frequently below the timberline.

For slope evolution, the important period is the main phase of ice retreat, which coincides with global warming and expansion of forests during Late Glacial and early Holocene. This phase is associated with the relaxation of slopes by release confining pressure connected also with isostatic rebound leading to slope retreat by formation of great slope failures and talus deposits at the base of chutes cutting the rocky cliffs (Ballantyne, 2002). Such features, among them great landslides and rockfalls have been described from the Alps (Panizza & *alii*, 1997), Pyrenees (Maya & *alii*, 1997) and Scandinavian Mts. (Rapp, 1960). Many of these features closer to the permafrost limit have been transformed to rockglaciers (Kaszowski, 1984; Sas-Walling, 2003). Parallel to these geomorphic responses to deglaciation have been the activation of debris flows monitored among others in the Tatra Mts. (Kotarba, 1988) and the entrenching of river channels (Fort, 2003).

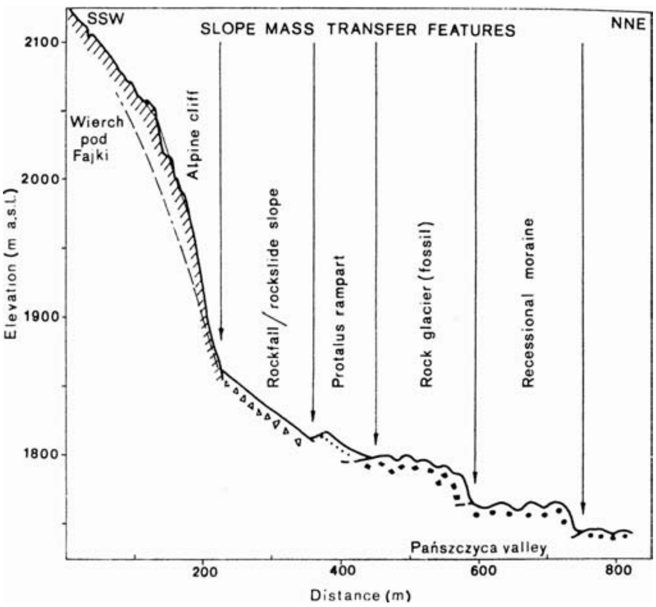

FIG. 8 - Slope located in one vertical belt (cryonival) earlier transformed by glacial erosion with wide paraglacial belt, Tatra Mts (after Kotarba & *alii*, 1987).

Especially active are the valley sectors of frequent ice advances and retreats during Neoholocene reflecting climate oscillation. Watanabe & *alii* (1998) have described such examples from Langtang Himal in Nepal, dating four generation of fans connected with supraglacial processes separated by entrenching of river channels during phases of glacier retreat. Similar processes of rapid recession and erosion are now monitored in detail in many river valleys in connection with present global warming (Kotlyakov, 2007; Owen & Sharma, 1998).

SHIFT OF VERTICAL ZONES DURING LAST GLACIAL-INTERGLACIAL CYCLE

The last cold stage was characterised by lowering of all climatic-vegetation vertical zones by about 600-1200 m including the position of the snowline, permafrost limit and timberline (Frenzel & *alii*, 1982; Starkel, 1977, 1998). The result was that slopes of lower mountains and foothills in the temperate zone have entered in the periglacial-cryonival belt with continuous permafrost and were modelled by cryogenic processes and solifluction (Poser, 1977; Starkel, 1968). A good indicator of that lowering are fossilised cryoplanation terraces on ridges in central Europe and in Mongolian mountains (Kowalkowski & Starkel, 1984 - figg. 9-10). The more active processes like debris flows were restricted to S-facing slopes, still preserved in a fossil form (Kowalkowski & *alii*, 1977).

At the same time, above the snowline valley glaciers developed and in some mountain ranges of the boreal zone and over the Tibetan Plateau even ice sheets or smaller ice caps (Shi & *alii*, 1992). In case of the Alps, long valleys were completely filled by glaciers, which at the mountain forelands extended as piedmont ice platforms. Similar piedmont glaciers existed at the northern part of the Tibetan highland in the Qilian Shan and Tanggula Shan (Hövermann & Lehmkuhl, 1993). The supraglacial part of mountain slopes was restricted mainly to knife-edge nunataks.

The most pronounced change at the continental scale was the expansion of continuous permafrost towards west, typical now for arid continental climate of northern Asia (Frenzel & *alii*, 1982). In central Europe, whole mountain

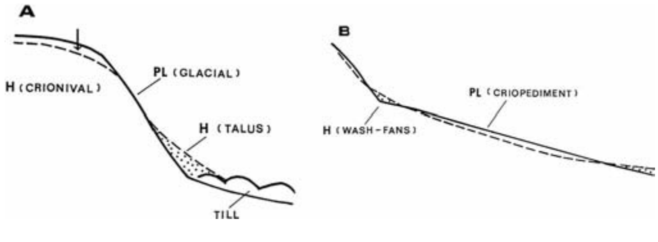

FIG. 9 - Change in the tendencies of slope evolution in Khangai Mts. - cold stage and Holocene (after Kowalkowski and Starkel 1984): a/ high mountain zone (2700-3500 m a.s.l.), b - arid submontane depressions (2000-2400 m a.s.l.). 1. primary profile (late Pleistocene), 2. trend of transformation during the Holocene, 3. direction of change.

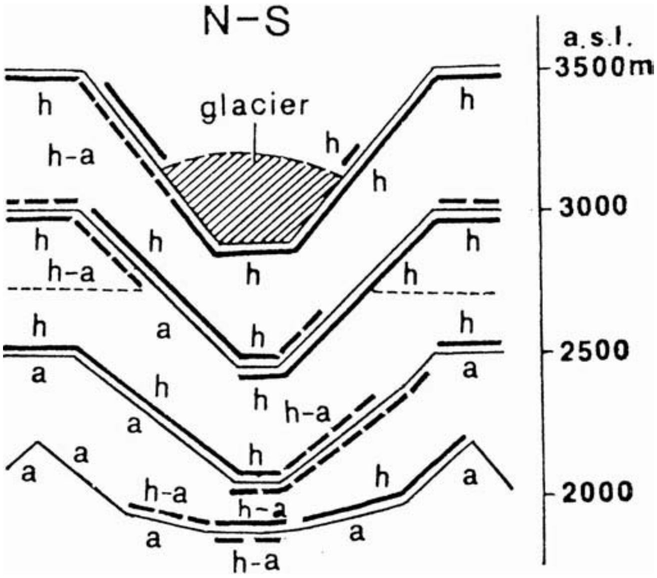

FIG. 10 - Difference in vertical zonality and asymmetry between last cold stage and Holocene in different altitudes of Khangai Mts. Above the profile the conditions during cold stage, below - in Holocene, thick line - continuous permafrost, striped line - discontinuous permafrost, h - dominance of processes of cryosemiarid belt, a - dominance of cryosemihumid belts, h-a transitional zone.

ranges were occupied by permafrost with tundra-steppe vegetation (Starkel, 1977). On the Balkan peninsula above about 1000 m as.l. the cryosemihumid zone with tundra vegetation and solifluction and below about 500 m as.l. the cryosemiarid zone with steppe vegetation and loess deposition were separated by a narrow forest-tundra zone with refuges of many tree species (Bottema, 1974; Starkel 1993 - fig. 11).

During 3000-4000 years of Late Glacial there followed a gradual warming and shift of climatic vegetation belts with trees expanding over former permafrost and over areas covered by glaciers. This shift reached up to 1000 m. During some phases, the rate of permafrost melting was very fast. This is indicated by rapid retreat of permafrost limit in Siberia (Baulin & *alii*, 1984; Nechaev, 2008). The deep percolation of water in the ground activitaed mass movements on the mountain slopes (Starkel, 1997; Panizza & *alii*, 1997). The extension of forest upslope was delayed after ice retreat by presence of free-face surfaces or coarse debris without soil cover. Still upper segments of slopes remained in the cryonival belt. The hills in Southern Sweden after ice sheet retreat and before occupation by forest at first were modelled by periglacial processes (Rapp, 1987). During the millennial episode of the Younger Dryas it was a distinct return to cooler and frosty conditions reflected in the reactivation of gelifluction, patterned grounds and over Scottish mountains in the lowering of snow line to 300 m (on the west side), 700 m asl. (on the east), leading to the formation of shallow ice cap (Sissons, 1977; Clapperton & Sugden, 1977).

After rapid warming about 11500 cal yrs. BP during the Holocene, followed the general stabilisation of morphodynamic vertical zones and the timberline reached the level similar to the present-day. The gradual rise of timberline in the Caucasus is marked by steps built of debris fans (Muratow, 1983). The former cryonival belt in most part of profile has been occupied by forest and gradually the soil profile was developed. The main role in the Holocene is played by chemical processes. The physical processes are restricted to mass movements, piping and linear erosion, all connected with water infiltrating in the ground. In valley heads, the entrenching river channels have lowered the base of slopes (Starkel, 1959).

During the Holocene, the second order climatic fluctuations cause the shifting up and down of the limit between forest and cryonival belts. This is documented by dating of patterned ground, solifluction lobes in lower elevation (Furrer, 1978; Gamper, 1993; Matthews, 1993; Rączkowska & *alii*, 2007) and alternatively above the treeline by presence of subfossil karren typical for forest zone (Kotarba & Starkel, 1972) and by higher extent of podsol soils (Adamczyk, 1962).

Coincidental with these changes are the fluctuations of glacier advances and retreats (Grove, 1997), lake level changes and sediments deposited in them (Magny, 1993; Kotarba & Baumgart-Kotarba, 1997) reflecting the phases of debris flow activity. In the vertical scale of mountain slopes, these fluctuations of forest upper limit or of periglacial phenomena do not exceed 250-350 m in elevation (Veit, 1993).

In the Holocene evolution of Eurasian mountain vertical zones, an important role is played by different sequence of temperature and humidity changes in the S-N transect (Starkel, 1998 - fig. 12).

In the lower latitudes the early rise of precipitation in Lateglacial and early Holocene preceded the rise of temperature. This caused a very slow rise of snow line (with still expanding glaciers) and relatively fast rise of timberline due to increase of monsoon rainfalls (Kutzbach, 1992). In higher latitudes, the global warming with still

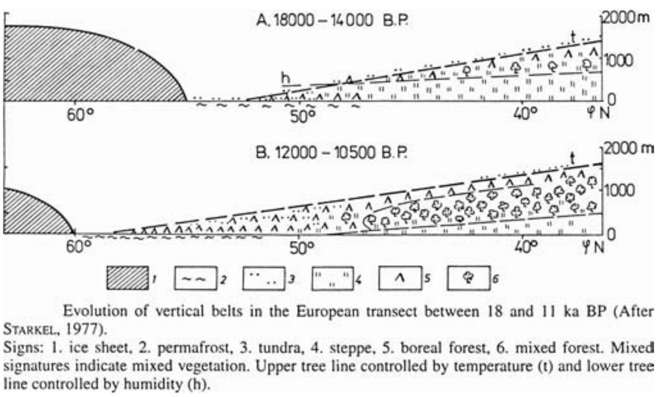

FIG. 11 - Evolution of vertical belts in European transect between 18 and 11 ka BP (after Starkel 1977). 1. position of ice sheets, 2. permafrost, 3. tundra, 4. steppe, 5. boreal forest, 6. mixed forest; mixed signatures indicate mixed ecosystems, t - upper tree line controlled by temperature, h - lower tree line controlled by humidity.

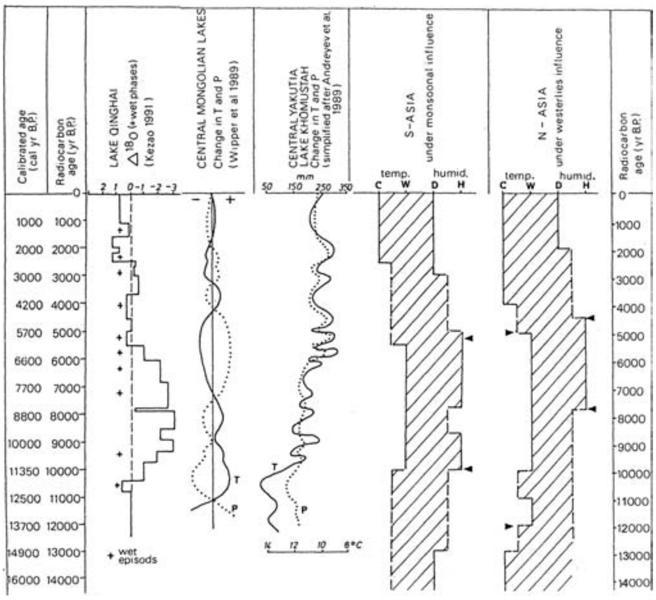

FIG. 12 - Changes of temperature and humidity (precipitation) in late glacial and Holocene in southern and northern part of Asian transect (arrows show time of distinct changes) - after Starkel 1998.

weak activity of westerlies (till 9500-9000 cal yrs BP) caused fast retreat of continuos permafrost and extension of boreal trees upslope (Khotinsky, 1984; Nechaev, 2008). Therefore the steppe vegetation in early Holocene still occupied the intramontane depression in Greece (Bottema, 1974 - fig. 11). and Mediterranean woodlands expanded very late during the Atlantic phase (Beug, 1982).

Starting from about 5 ka BP the general tendency to cooling in Asia is observed as well as gradual aridization (Bryson, 1997). It is expressed in the lowering of permafrost belts and limit of cryonival processes, marked on the Tibetan Plateau in covering of peaty soils by solifluction lobes in the altitude between 4000 and 5000 m asl. (Starkel, 1998).

The type and intensity of processes typical for climatic-vegetation zones have been changed in last millennia by deforestation and overgrazing, which caused the reactivation not only of cryogenic processes but also of slope wash, debris flows and deflation, especially active in zone of lowered timberline (Starkel, 1987).

SEQUENCE OF VERTICAL ZONES IN ONE SLOPE PROFILE

The difference between the last cold stage and the Holocene in vertical zonation of mountain slopes depends on relief energy, maturity of relief (slope profile), stability of base level and lithology of substratum. The type and intensity of processes changes with elevation, therefore we can talk on slopes which embrace one, two, three or even four vertical zones.

The «one vertical belt» slopes are characteristic for lower slopes (up to 300-500 meters) which are frequent in foothill areas with mature relief or in hanging, not rejuvenated, headwater-valley sections and have relatively stable base level. In the temperate zone these are located at present in the forest belt, dominated by chemical processes, only locally modelled by gullies, piping and landslides. During the cold stage these slopes have been located in the periglacial zone with permafrost and frequently developed colluvial glacis or cryopediments. Depending on bedrock type these may still preserve inherited block fields (on the quartzite hills) and convex-concave shape (Czudek & Demek, 1973; Starkel, 1968).

Other «one-belt» slopes are close to present timberline or above it in valley heads of highest parts of various other mountain groups of central and south-eastern Europe glaciated in the Pleistocene. The free-face slopes of former glacial cirques and troughs are now modelled by cryonival processes and their stable base level facilitate the development of taluses and alluvial sectors of slopes. In the Tatra, first Klimaszewski (1971a) and later, in detail studies, Kotarba (1984, 1988) have shown the mechanism of gradual retreat of free-face slopes and growing of talus-alluvial section.

But for the whole temperate zone of Eurasia where the relief energy fluctuates between 400-1000 m most typical are the slopes located in two vertical belts, both during cold stage as well during the Holocene. In the flysch Carpathians the lower parts of slopes during cold stage were modelled by gelifluction and the upper steeper rocky slopes underwent transformation to cryoplanation terraces (Klimaszewski, 1971b; Zietara, 2002). At present the slope is mainly within whole length in the forest belt (fig. 13).

At higher elevations the bi-partial slopes during cold stage were located in cryonival belt and their upper parts above snowline where the glaciers were formed and later sloped down valley. At present these slopes are totally above forest zone in the altitude between 1500 and 2500 m asl. and the free-face segments with intensive frost weathering supply talus of coarse debris (Kotarba, 1988; Rączkowska, 2007 - cf. fig. 1).

The higher slopes (1000-2000 m and more) occupy 2-3 vertical zones and usually are passing from the glacial belt across two-partial cryonival belts down to the forest zone. Depending on rock resistance and the rate of physical weathering, the steep free-face segments reach various heights; for instance, in Hindukush pass 1000 m (Kaszowski, 1984) and the slope of Nanga Parbat even 4000 m (Shroder & Bishop, 1998). The aggradation at their base depends not only on the supply from upper section but also on the transfer of debris by rivers.

All these factors, especially altitudinal position of the free-face section in relation to zonal intensity of processes, decide which part of slope is the most productive: upper, middle or lower one (fig. 13).

The longest slopes starting above snowline and ending in the forest zone are most active in the middle part

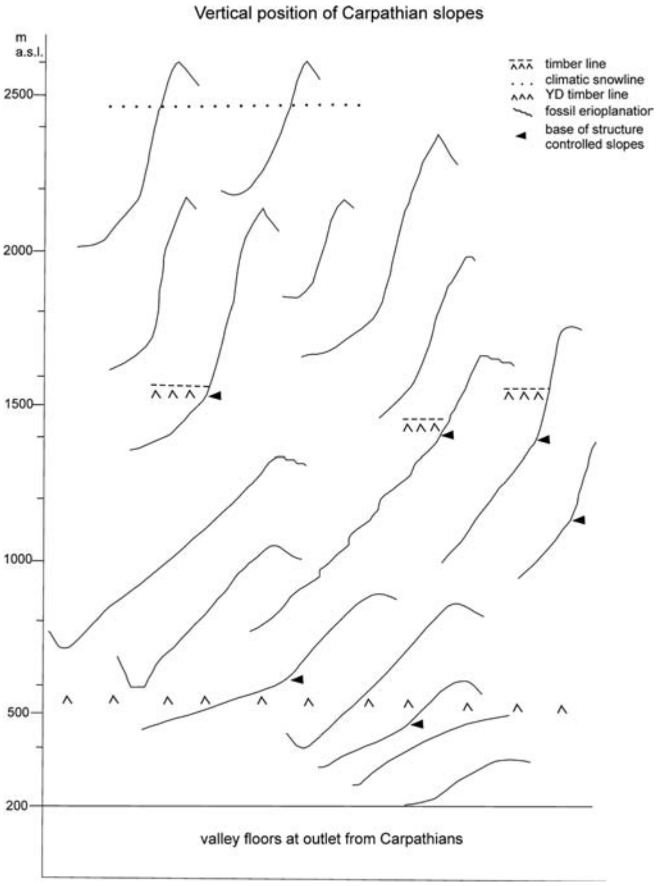

FIG. 13 - Location of most frequent slopes in climatic vegetational vertical zones of the Polish Carpathians. 1. present timber line, 2. present snowline, 3. Younger Dryas timber line, 4. fossil cryoplanation, 5. slope break on the structure controlled slope.

by passing the cryonival zone. Those ending in cryonival have a most active middle or lower part. Slopes starting in cryonival and continuing in the forest zone have a more productive upper part. Slopes in mountains of Central Asia with a narrow forest zone separating criosemihumid from criosemiarid have a less active middle section of slope (fig. 5, 10). Finally the slopes undermined by entrenching river channels show the highest intensity at their base.

The production of debris over slopes supplied to their base and later transferred by the river cause differenciated tendencies in slope evolution, going in opposite directions: from slope retreat and lowering of base level to progressing aggradation by talus, rockglaciers etc.

CONCLUSIONS

The analysis of climatic-vegetation vertical zones in the Eurasian mountains and their reflection in slope evolution during last climatic cycle in the Quaternary presents a great diversity connected with a range of factors. Some of these factors seem to be independent of change in temperature with elevation and long term climatic fluctuations, like tectonic uplift, resistance of substratum or formation and melting of mountain glaciers. But in reality, all of them are connected as it is shown on the simple model of mountain geoecosystems (fig. 14). Tectonic uplift creates the background for existence of mountain ranges. The valley glaciers disturb the regular vertical zonality and are in fact the product of highest mountain belt above the snowline.

Discussion on changes of vertical climatic zonality in the Eurasian mountains reflected in the diversity and evolution of slopes helps to formulate following conclusions:

1. Climatic-vegetation vertical zones shifting between cold and warm stages of Quaternary determine the rate and direction of slope evolution. Under the forest cover the forms and deposits from previous cold stage are preserved.
2. In the Eurasian mountains, beside vertical displacement of morphoclimatic zones during cold stages, a more pronounced role is played by the expansion in western direction of continentality with permafrost.
3. Vertical zones of present-day geomorphic processes and their limits are frequently adapted to inherited features of relief.
4. Continuous high rates of uplift and deepening of valleys cause an undermining of slopes and large-scale slope failures connected with extreme events. This trend may play a leading role in evolution of young mountains, more important than the climatically controlled secular processes characteristic of various climatic zones.
5. The mountain glaciers are superimposed on vertical zones being simultaneously also their product. During advances the glacial tongues separate valley sides into supraglacial, paraglacial and subglacial sectors.
6. In the long-term perspective the Quaternary uplift of young mountain chains up to 1-2 km means the gradual shift of erosional mountain relief to higher climatic zones and progressive adaptation of fluvial landscape to cryonival and glacial environment.

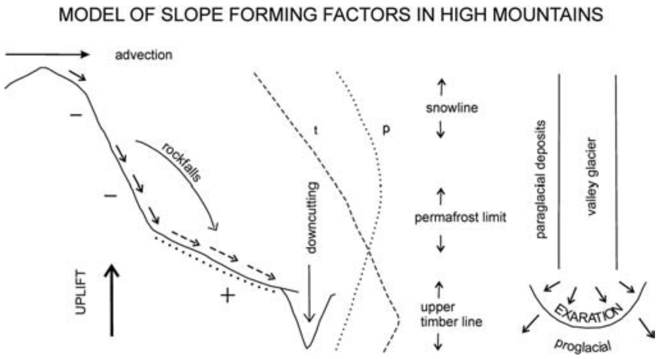

FIG. 14 - Simplified model showing main factors and processes transforming the high mountain slopes.

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