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The first stages of erosion by ice sheets: Evidence from central Europe

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ABSTRACT

In almost all former and presently glaciated areas of the world, glaciers have modified or even transformed pre-glacial terrain during the many cold stages of the Pleistocene. In consequence, the early stages in the development of glacial landscapes have been overprinted or erased by later phases of erosion.

The Sudetes in central Europe provide exceptional opportunities to examine the inception of glacial erosion. Evidence of a long geomorphic evolution before glaciation, with the development of etch surfaces, deep weathering covers and the preservation of Neogene kaolinitic sediments, provides a direct analogue to other lowland crystalline terrains as existed immediately prior to Pleistocene glaciation. Large granite intrusions exist in the Sudetes and its Foreland that support hills that range in size from small tors to large domes. These terrains experienced only thin ice cover for short periods when the Scandinavian ice sheet reached its Pleistocene maximum limits in Marine Isotope Stage (MIS) 12 (Elsterian) at 440–430 ka and MIS 8 (Early Saalian) at 250–240 ka. In each of four study areas beyond, at and within the limits of glaciation we have mapped Glacial Erosion Indicators, glacial landform assemblages that indicate progressive glacial modification of the pre-glacial granite terrain.

We find that glacial erosion increases with distance from the ice margin, due to greater ice thickness and longer ice cover, but has had only limited impact. On hills, regolith was stripped by moving ice, tors were demolished, and blocks were entrained. However, indicators of more advanced glacial erosion, such as lee-side cliffs and glacial streamlining, are absent, even from granite domes that lay beneath ~500 m of ice. The survival of tors beneath ice cover and the removal of tor superstructure by glacial erosion are confirmed. The presence of glacially-modified tors in areas covered only by Elsterian ice implies that the tors existed before 440 ka. Moreover, the contrast between blockfield-covered slopes found above Elsterian and Saalian glacial trimlines on summits of granite, gabbro and basalt hills in the Sudetic Foreland and glacially-stripped surfaces at lower elevations implies only limited regeneration of blockfields since 430 and 240 ka. The first stages of roche moutonnée formation are recognised in the exhumation of tor stumps but development of lee-side steps remains very limited. Granite domes retain a pre-glacial morphology and local examples of hill asymmetry are determined by structural, rather than glaciological controls. Development of roches moutonnées and asymmetric and streamlined hills by glacial erosion has not occurred in the Sudetes despite the passage of largely warm-based ice sheets 80–500 m thick that persisted for 5–20 ka.

1. Introduction

Whilst landscapes of little or no glacial erosion have been recognised in many formerly glaciated areas (see summary in Benn and Evans, 1998), the changes brought about at the inception of glacial erosion have only recently begun to be identified (André, 2004; Hall and Sugden, 2007). This limited understanding reflects the present condition of almost all former and presently glaciated areas of the world, where pre-glacial terrain has been modified or transformed by glacial erosion during the many cold stages of the Pleistocene. In consequence, landforms created early in the development of glacial landscapes have been overprinted or erased by later phases of erosion and so are difficult to decipher. Yet understanding the geomorphic changes that take place at the inception of glacial erosion is important for several reasons. Differences between the weathered and permeable materials of the pre-glacial landscape and the hard rock or drift-covered substrates more typical of glacier beds in the Late Pleistocene may have led to important differences in ice sheet behaviour (Clark and Pollard, 1998). Reliable recognition of relict landforms is required in order to identify patterns of erosion in glaciated landscapes (Hall and Sugden, 1987) and for former glacier basal thermal regimes to be reconstructed (Kleman and Stroeven, 1997; Stroeven et al., 2002; Harbor et al., 2006). Models of the progressive development of landforms of glacial erosion, including roches moutonnées (Jahns, 1943; Sugden et al., 1992; Glasser, 2002) and domes (Matthes, 1930; Ebert and Hätterstrand, 2010) also need to be tested against field

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evidence in order to establish the significance of inheritance in the formation of glacial landforms (Lidmar-Bergström, 1997).

Previous attempts to examine the inception of glacial erosion have relied on theoretical models (Harbor, 1995; MacGregor et al., 2000), on comparisons of unglaciated and glaciated terrain (Feininger, 1971; Hall, 1985; Lidmar-Bergström, 1996; Johannsson et al., 2001) and on space-time substitutions which place large glacial landforms in sequences of progressive morphological development (Sugden et al., 1992; Hall and Phillips, 2006). We adopt a new approach to this problem by examining the morphology of granite hills on the northern footslopes or Foreland of the Sudetic Mountains (Fig. 1) of central Europe. The Sudetic landscape provides an analogue for the geomorphology of crystalline terrains as they existed before Pleistocene glaciations, with extensive deep weathering, pre-glacial sediments and hill forms at different scales. Whilst the Scandinavian ice sheet advanced across the north Polish plain on many occasions, it reached maximum limits as far south as the Sudetic Foreland only in two or three phases during the Middle Pleistocene (Macoun and Kralik, 1995; Badura and Przybylski 1998; Marks, 2004). Hence the study area provides exceptional opportunities to examine the first stages of glacial erosion because the hills have experienced only brief (5–20 ka) periods of glaciation beneath the margins of ice sheets of restricted thickness (80–500 m). Our approach to the problem is geomorphological. We map the morphology of four basement (mainly granite, also gneiss and schist) terrains at different distances from the former ice margins in terms of Glacial Erosion Indicators (GEI), i.e. glacial landform assemblages that indicate progressive glacial modification of the pre-glacial terrain.

Our approach provides a fresh perspective on the development of the landscapes of the Sudetes and its Foreland during the Pleistocene, where previously little consideration has been given to the impact of glacial erosion (Demek, 1976). Changes at the early stages of glacial erosion of granite hills from the scale of tors to domes are identified. Spatial patterns in glacial erosion near the margins of the Scandinavian Ice Sheet provide perspective on the ice thicknesses (Glasser, 1995) and timescales (Briner and Swanson, 1998; Kleman et al., 2008) necessary to induce further glacial erosion and the scouring and streamlining of resistant bedrock. The study demonstrates that detailed geomorphological mapping can identify patterns of pre-glacial inheritance and glacial erosion, which can be further tested against numerical modelling or inventories of cosmogenic isotopes.

2. Study area

The Sudetes mountain range is almost 300 km long and up to 100 km wide and forms the north-eastern rim of the Bohemian Massif in Central Europe (Fig. 1). This study examines the inception of glacial erosion on granite hills and their metamorphic surroundings in four areas of Poland and the Czech Republic: the Jelenia Góra basin in the west Sudetes, and the Žulova massif, the Strzelin massif, and the Strzegom-Sobótka massif in the Sudetic Foreland (Fig. 1).

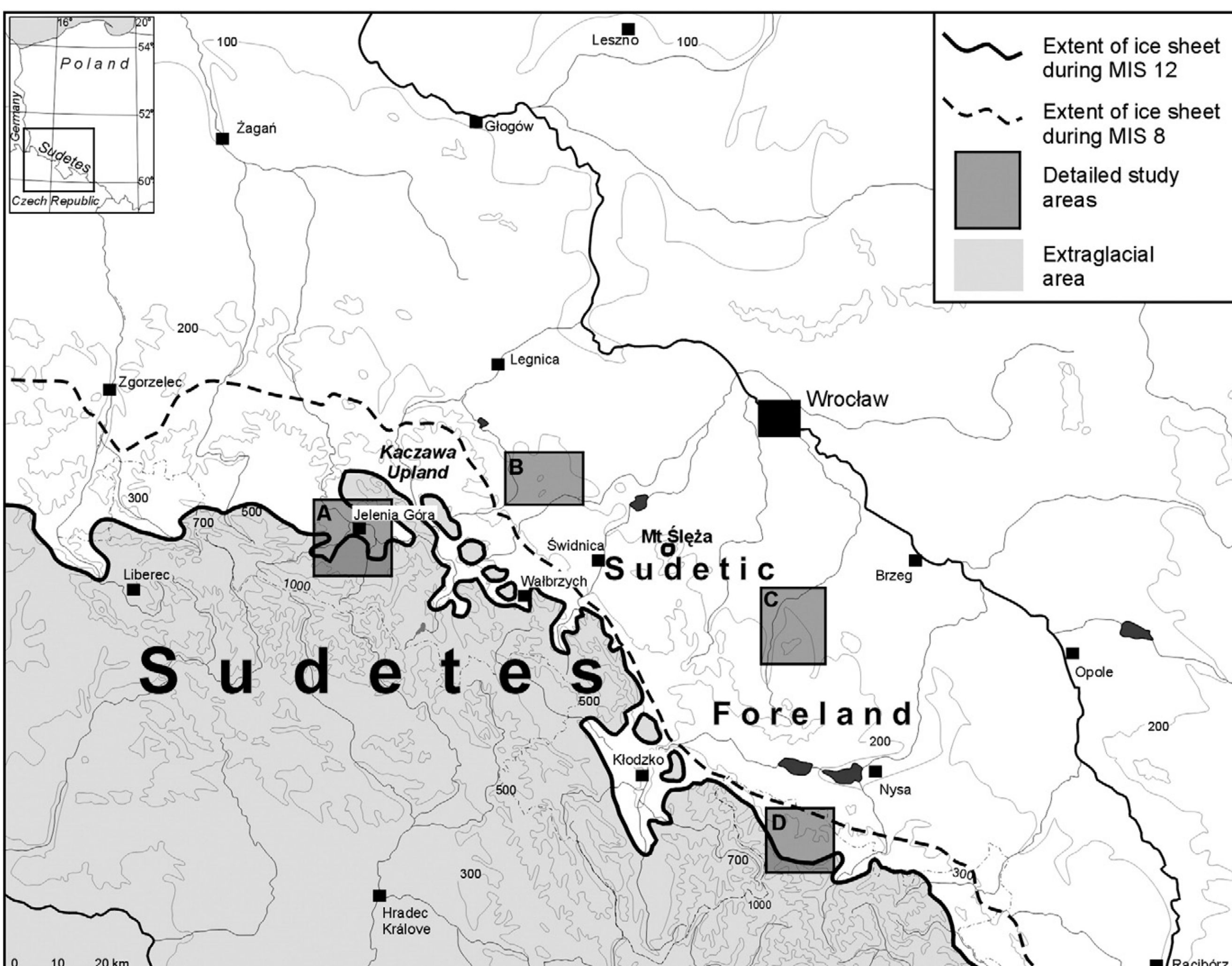

Fig. 1. Location map of the Sudetes, its Foreland, and the Polish plain.

2.1. Geology and tectonics of the Sudetes

The geological history of the Sudetes is highly complex and involves many stages of magmatism, sedimentation and deformation, spanning parts of the Precambrian and the entire Phanerozoic (Aleksandrowski and Mazur, 2002). The ultimate consolidation of the structure took place during the Variscan orogeny in the Devonian and Carboniferous, with emplacement of numerous granite intrusions (Mazur et al., 2007). Long-lasting, although intermittent terrestrial and marine deposition followed until the end of the Cretaceous, whereas the Palaeogene is considered as the protracted period of deep weathering, evolution of hilly to inselberg relief, or planation in various settings.

In the last 20 million years, the Sudetes have been subject to neotectonic uplift and subsidence (Zuchiewicz et al., 2007). The most evident landforms of tectonic origin are straight mountain fronts, among which the one associated with the Sudetic Marginal Fault is the most prominent (Krzyszkowski et al. 1995; Badura et al., 2007). Movement of the Sudetic Marginal Fault, probably starting in the late Oligocene/early Miocene and continuing today (Badura et al. 2007, Štěpančíková et al. 2008), has divided the area into the relatively upthrown block of the Sudetes proper and the subsided block of the Sudetic Foreland. As a result, denudation accelerated in the elevated area, whilst widespread accumulation commenced on the Foreland, covering the deeply weathered Palaeogene surface.

2.2. Preglacial basement landscapes

The Sudeten topography is complex and juxtaposes mountain ranges, plateaux, and dissected massifs with troughs and intramontane basins. Relative relief is high, with basin floors at altitudes of 300–500 m overlooked by mountain summits at >1000 m. North of the Sudetes range and separated from it by a straight mountain front lies the Sudetic Foreland. The Foreland has a hilly topography, with numerous bedrock inselbergs 50–500 m high rising above a rolling surface underlain by Neogene and Quaternary deposits. Palaeogene deep weathering covers are widely preserved in the Sudetic Foreland and form part of a residual relief that consists of plains, inselbergs, and ridge and valley topography with a height range of up to 600 m.

As inferred from the widespread presence of pre-Quaternary deposits around bedrock elevations in the Sudetic Foreland, before the onset of glaciation, the positive relief elements of the Sudetes and its Foreland must have been similar to that of today. Bedrock on low ground was often masked by Neogene deposits and generally deeply weathered, with remnants of Palaeogene kaolinitic saprolites many tens of metres thick in the Foreland and grus-type saprolites developed on granite during the late Neogene found throughout the lower parts of the Sudetes. The relief was hilly, with rock outcrops confined largely to the upper slopes of ridges, inselbergs, and tors that rose from surrounding structurally-guided basin and valley floors. The emergence of these hills was a result of differential weathering and erosion sustained over periods of 10^7 yr. The Cenozoic etch landscape of the Sudetic Foreland has many parallels with other deeply weathered crystalline terrains found beyond the limits of glaciation in Europe, in southern part of the Bohemian Massif (Ivan and Kirchner, 1994; Roetzel and Roštinský, 2005), France (Klein, 1997) and Britain (Coque-Delhuille, 1991). Within the limits of Pleistocene glaciations many features survive in zones of limited glacial erosion that are also integral to the Sudetic landscape, such as kaolins (Kaye, 1967), grus (Migoń and Lidmar-Bergström, 2001), pre-glacial deposits (Hall, 1985), tors (Hall and Sugden, 2007) and inselbergs (Ebert and Hätterstrand, 2010). The Sudetic Foreland can be seen as one variant of the diverse reliefs that existed before Pleistocene glaciations on crystalline platforms.

2.3. Glacial history of the study area

The Scandinavian ice sheet reached its maximum southern extent in the Sudetes (Badura and Przybylski, 1998). Marginal ice lobes moved along valley tracts, across low altitude passes, and towards intramontane basins, where proglacial lakes formed (Schwarzbach, 1942; Jahn, 1960; Kowalska, 2007).

The vertical ice limit at the Pleistocene Glacial Maximum in MIS 12 is uncertain. Near ice margins in the Sudetes till-derived diamictons and scattered erratics are found on passes reaching 500 m asl. (Badura and Przybylski, 1998). To the north, prominent isolated hills, including Mt. Ślęża, stand within the regional ice sheet limit (Fig. 1). Angular tors, blockfields, and block slopes that are confined to the upper slopes of such hills have been inferred to stand above the former ice sheet surface (Szczepankiewicz, 1958; Martini, 1969; Präger, 1987; Żurawek and Migoń, 1999). According to this nunatak interpretation, ice thickness was 300–400 m over the Sudetic Foreland. In eastern Germany, however, ice thickness 50 km north from the Elsterian maximum ice front was ~500 m (Eissman, 2002), requiring basal shear stresses of ~20 kPa. Projection of simple ice profiles north from limits in the Sudetes and its Foreland using this value and the method of Nye (1969) implies that the MIS 12 ice sheet covered almost all Foreland summits and reached a thickness of 500 m over the Strzegom Hills (Fig. 2). The greater ice thickness suggests that alternative interpretation of the trimlines as marking englacial thermal boundaries (Ballantyne, 1997) is more likely, with preservation of delicate features on summits beneath cold-based patches. Whilst the nature of former subglacial environments in the Sudetic Foreland remains poorly understood, neighbouring areas of east Germany and the foreland of westernmost Sudetes and Erzgebirge provide analogues. At lower elevations, the rapid speed of ice advance (Junge, 1998), the large volumes of meltwater required for the excavation of sub-marginal tunnel valleys (Ehlers and Linke, 1989) and the planar nature of unconformities beneath tills (Piotrowski et al., 2001) together indicate widespread sliding beneath the warm-based marginal zones of the Elsterian ice sheet.

The timing of the ice sheet advances is poorly constrained. From the 1950s to early 1990s, two ice advances into the Sudetes, of almost identical extent, were suggested and correlated with the Elsterian glaciation (equivalent to MIS 12) and the older Saalian glaciation (equivalent to MIS 8) (e.g. Jahn 1960, Szczepankiewicz 1989). However, unequivocal geological or geomorphological evidence for two separate ice sheet advances into the Sudetes is lacking and recent interpretations link all glacial deposits in the Sudetes, perhaps except the very marginal, low-lying parts, with the Elsterian ice sheet (Badura and Przybylski, 1998).

To the north in the Sudetic Foreland, in contrast, two or three superimposed till horizons have been found intercalated with non-glacial sediments. These sequences have been interpreted as evidence for two ice advances during MIS 12, and a single ice advance during MIS 8 (Krzyszkowski and Czech 1995; Badura et al., 1998). The MIS 8 ice sheet was much thinner in the Sudetic Foreland than its MIS 12 precursors and therefore unable to cross the morphological barrier of the mountain front. The geographical limit of the MIS 8 ice sheet is suggested to broadly coincide with the 300 m asl contour line (Badura and Przybylski, 1998). The youngest tills may relate to MIS 6 glaciation (Marks, 2004), but this has yet to be confirmed by dating.

This latter reconstruction has important implications for estimating the timing and duration of ice cover in the study areas. The southernmost areas of the Jelenia Góra Basin and the Žulova Hilly Land would have been glaciated only during MIS 12. In contrast, the northern areas of the Strzegom Hills and Strzelin Hills were ice-covered during MISs 12 and 8. Although the total duration of these glacial phases is >100 ka, marine isotope curves indicate that the

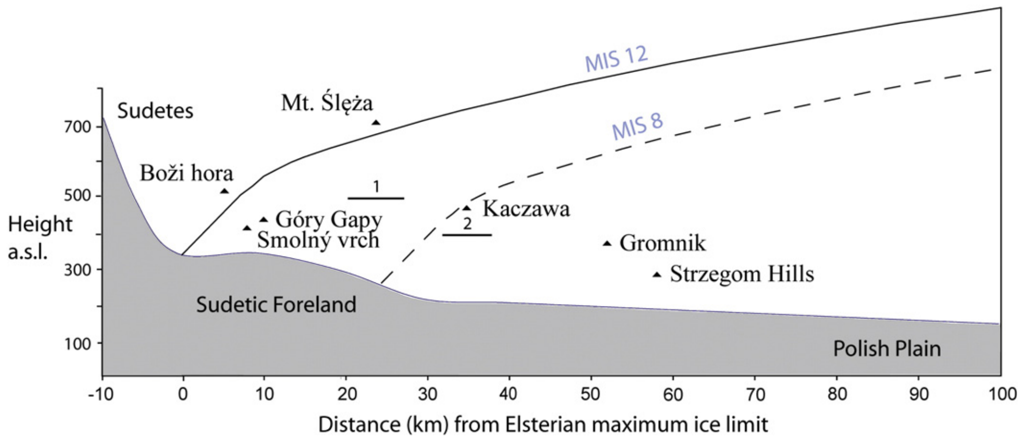

Fig. 2. Projected ice sheet profiles for the Elsterian and Saalian (dashed line) ice sheets. Inferred glacial trim lines at Mt. Ślęża (1) (Żurawek and Migoń, 1999) and in the Kaczawa Upland (2) (Migoń et al., 2002) fall below the reconstructed vertical limit of the Elsterian ice sheet.

duration of individual glacial maxima was <10 ka (Porter, 1989). This accords with evidence from varve counts in ice-dammed lakes at the Elsterian and Saalian maximum limits, where ice dams generally persisted for <5 ka (Schwarzbach, 1942). Thus, periods of actual glacial erosion of bedrock hills may have been very short, approximately 10 ka each.

3. Methods

This study focuses on the systematic search for morphological features regarded as indicative of the inception and progress of glacial erosion in granite terrain. These Glacial Erosion Indicators (GEIs) include, in approximate order of increasing glacial modification, the following:

(i) stripping of regolith from and transport of boulders across low angle surfaces (Kleman, 1994),
(ii) progressive removal and transport of large blocks and superstructure from rock outcrops, including tors, in circumstances that preclude toppling, rock wall collapse and periglacial mass movement downslope (André, 2004; Hall and Phillips, 2006),
(iii) preferential development of small lee-side steps on small granite protuberances (Hall and Phillips, 2006),
(iv) streamlining of small hills and development of lee side cliffs, with the formation of roches moutonnées (Jahns, 1943; Lindström, 1988; André, 2001),
(v) development of large scale hill asymmetry, with stepped cliffs on the lee side of the hills (Sugden, et al., 1992) and
(vi) streamlining of large hills parallel to former ice flow (Rudberg, 1954; Ebert and Hätterstrand, 2010).

Only hill forms and their surroundings were considered in detail in this study of glacial erosion in the Sudetes because exposure in basins and valleys is often very poor because of covers of löess, periglacial slope deposits and, locally, glacial deposits. Dense forest cover on hills prevents air photo interpretation, and morphological mapping was carried out mainly on the ground. Several of the key areas are covered by 1:15, 000 orienteering maps that show all major rock outcrops. The Sudeten granites have been quarried for many centuries. The quarries provide frequent, extensive exposures that supply important information on the links between hill form, granite structure, and the distribution of regolith, including saprolites.

4. Field evidence and interpretation

4.1. Jelenia Góra Basin and its surroundings

4.1.1. Evidence

The Jelenia Góra Basin is one of the largest intramontane basins in the west Sudetes. The basin floor at 350–40 m asl is underlain by granite of Late Carboniferous age (Mazur et al., 2007) (Fig. 3A). Selective weathering and erosion has given rise to hilly topography, with a number of individual inselbergs, as much as 250 m high (Migoń, 1996, 1997) (Fig. 3B). Rock outcrops are frequent on some inselbergs, whereas other domes have smooth, regolith-covered slopes or only occasional tors at or near the summit. In between the hills, the terrain is rolling to hilly, with secondary basins, low elevations, scattered boulders, and tors. A mantle of grus, a granular, sandy saprolite developed on granite under temperate climates (Migoń and Lidmar-Bergström, 2001) is ubiquitous on the floor of the Jelenia Góra Basin and reaches thickness of 15 m (Fig. 4).

The general topography of the basin floor does not radically change across the MIS 12 ice limit, with clusters of inselbergs separated by basins and troughs typifying both the northern and the southern part of the basin; but tor shapes vary. Whereas towers and castle koppies are frequent in the southern part, more subdued tabular shapes dominate in the north. At an altitude of <380 m asl, angular or boulder tors are completely replaced by exposed plinths of granite (Fig. 3B and 4).

The upland terrain to the NW is built of Early Palaeozoic granites and gneisses (Oberc-Dziedzic, 2007) and consists of broad elevations, with few tors, separated by troughs and depressions. Slopes are less steep than within the granite relief of the Jelenia Góra Basin, except in the proximity of the Bóbr gorge. Here, the Bóbr river is incised into the upland terrain by 80–120 m; and within steep valley sides, angular tors, pinnacles, and rock ribs have developed (Traczyk, 2007). Further tors occur beyond the gorge, within gentler slopes leading to the summits of granite-gneiss elevations, which rise to 450–500 m asl (Fig. 5). Hillslope and summit tors vary from low elevations 2–3 m high to imposing, angular towers up to 10–12 m high, surrounded by blockfields (Fig. 4).

4.1.2. Interpretation

The maximum extent of MIS 12 ice in the Basin has been mapped on the basis of till deposits and erratic distribution (Berg, 1941; Fig. 3B). In this study, however, glacial erratic material has been identified on the summit of Gapy at 470 m asl (Fig. 5). This indicates

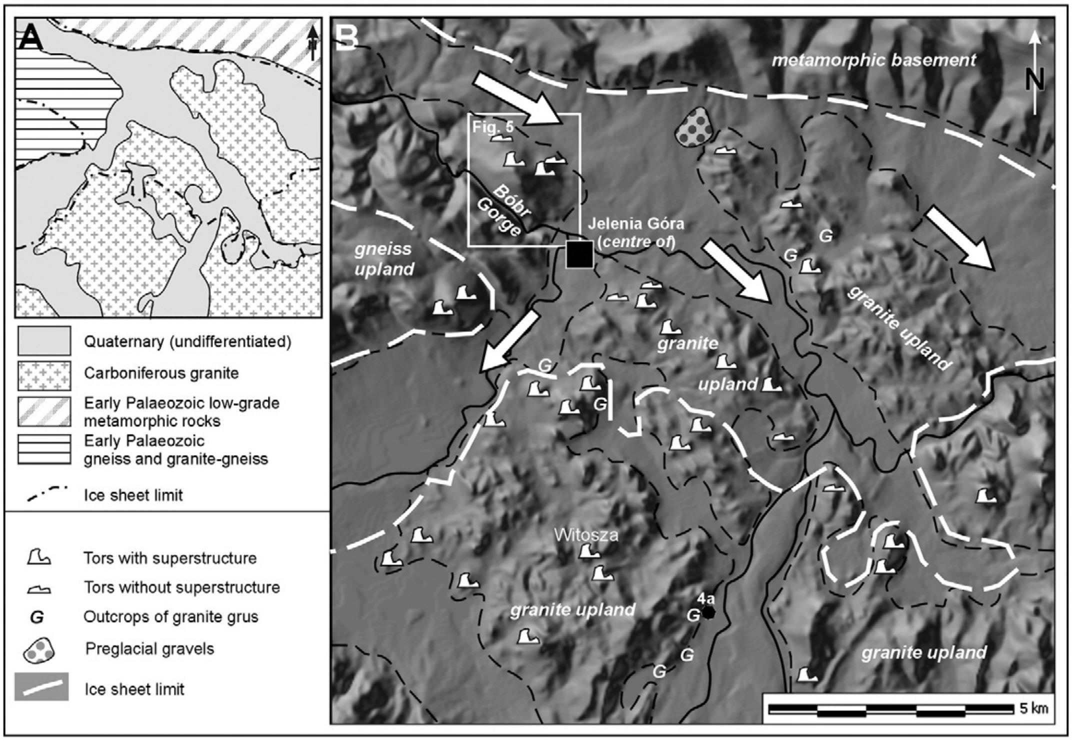

Fig. 3. Geology and geomorphology of the Jelenia Góra basin. (A) Geology. (B) Geomorphology. Dashed black line separates upland terrain built of solid rock from topographic depressions underlain by thick mantle of unconsolidated Neogene and Quaternary deposits.

that ice thickness over the northern part of the basin floor was > 130 m.

Glacial Erosion Indicators 1 and 2 are confined to low ground in the northern part of the basin. The absence or poor development of tors and the occurrence of displaced tor blocks east of Jelenia Góra are consistent with minor glacial modification. Although Zimmermann (1937) identified roches moutonnées in this area, we have been unable to identify forms with clear lee-side cliffs, and the rounded forms appear to be structurally-controlled tor stumps (Jahn, 1952). In the northern part of the Jelenia Góra Basin, delicate tor structures only occur higher than 390 m asl, i.e., 50–70 m above the valley floors, whereas in the south no such altitude constraint exists for tor occurrence (Fig. 3B). These tors, and the widespread grus on the basin floor, appear to have survived ice cover without glacial modification.

Glacial Erosion Indicators 3 and 4 appear to be absent. Domes studded with tors occur widely. Even at the locations most exposed to ice flow, little sign of glacial reshaping of hills is evident. Góry Gapy and adjacent hills lay directly in the path of ice entering the basin and some show a NW–SE orientation parallel to former ice

Fig. 4. Landforms of the Jelenia Góra basin. (i) Grus on the basin floor; (ii) pinnacle tor on Góry Gapy; (iii) subdued tor on Góry Gapy.

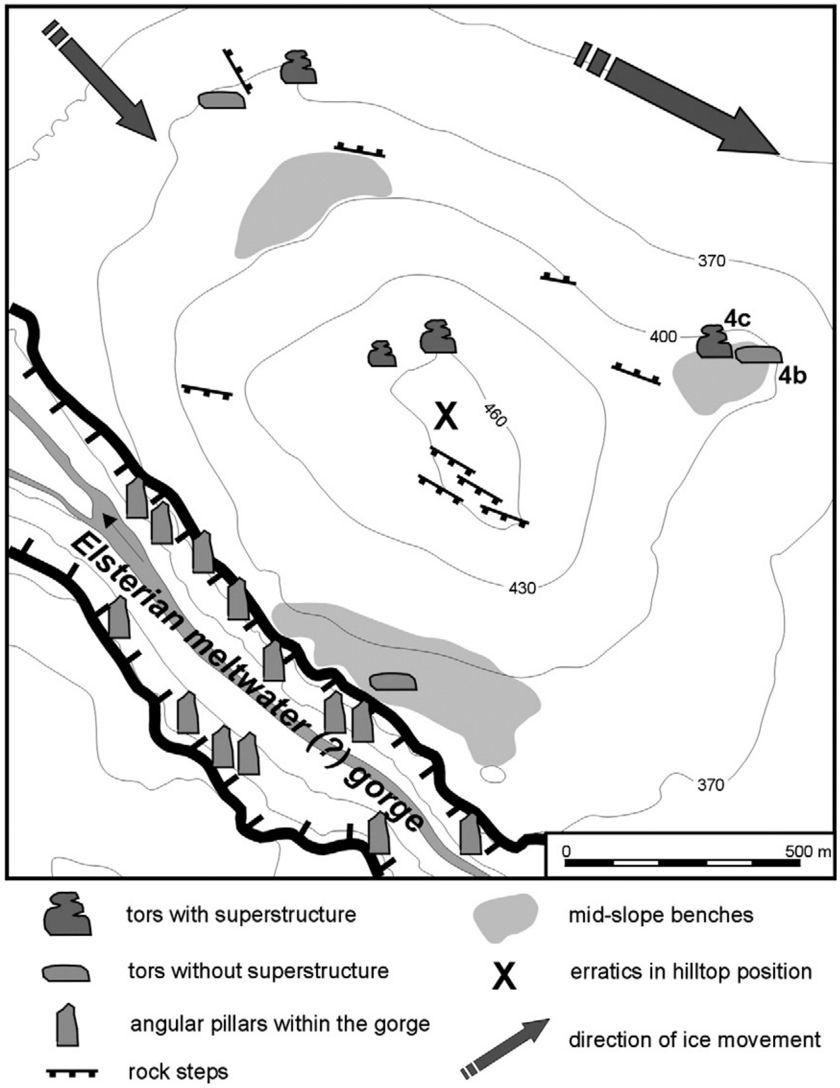

Fig. 5. Geomorphological sketch map of Góry Gapy.

flow. However, although the lee slopes of Gapy are slightly steeper than the stoss slopes, no lee side cliffs are evident. Moreover, angular tors up to 8 m high rise from the stoss-side slopes of Gapy, and smaller tors also occur on satellite hills. Granite outcrops indicate that the detail of the hill forms is controlled by sheet joints and that the orientation is mainly a reflection of vertical fracture patterns. Perhaps the main impact of glacial erosion has been in the deepening of the Bóbr gorge by meltwater, with subsequent removal of grus and the emergence of pinnacles over post-Elsterian time (Traczyk, 2007).

4.2. Žulova Hilly Land

4.2.1. Evidence

The Žulova Hilly Land (Fig. 1) lies immediately north of the topographic scarp of the Sudetes. The relief is developed in Late Carboniferous to Early Permian granite and granodiorite (Zachovalová et al., 2002) (Fig. 6A). In its northernmost part the intrusion is covered by thick Neogene and Pleistocene sediments, whereas kaolinitic weathering up to 50 m thick occurs along the northern edge of massif, occasionally cropping out to the surface (Kužvart, 1965; Gabriel et al., 1982; Kościówko, 1982) (Fig. 6B).

The Hilly Land includes more than 30 isolated hills, rising 50–150 m above the adjacent rolling surfaces (Demek, 1964, 1976; Ivan, 1983). The inselbergs typically are domes, and several display spectacular rock cliffs and are surmounted by tors. Other low domes (Fig. 7) occur in the northern part of the Hilly Land (Štěpančíková and Rowberry, 2008).

Smolný vrch (404 m) is a prominent dome about 60 m high, elongated in the SW–NE direction, and distinctively asymmetrical in profile (Fig. 8). The NE-facing side is gently inclined, with scattered boulder piles and small tors. To the SW, the hill terminates in cliffs up to 15 m high, separated by deep joint-aligned clefts, below which a moderately inclined slope occurs. The hilltop part of Borový vrch (487 m) shows a similar pattern, with a smooth surface facing NE and a tor-studded SW-facing slope and, to the south, the Píšťala dome (447 m asl) is also surmounted by a tor (Štěpančíková and Rowberry, 2008). All three dome summits carry a remarkable array of microforms, including weathering pits, tafoni and flutes (Demek, 1964) (Fig. 7). In contrast, low-lying rock surfaces in the Žulova area carry only weakly developed microforms.

The lower surface at 300–380 m from which the inselbergs rise is often rough, with a multitude of low elevations, oval or elongated in plan, and boulder-covered depressions (Demek, 1976) (Fig. 8). Small tors, with and without superstructure, occur around the base of Smolný vrch (Fig. 8). Open joints in tors SE of Kani hora contain glacial erratics (Štěpančíková and Rowberry, 2008). Tors also occur in the high ground south of the limits of glaciation. Sharp changes of

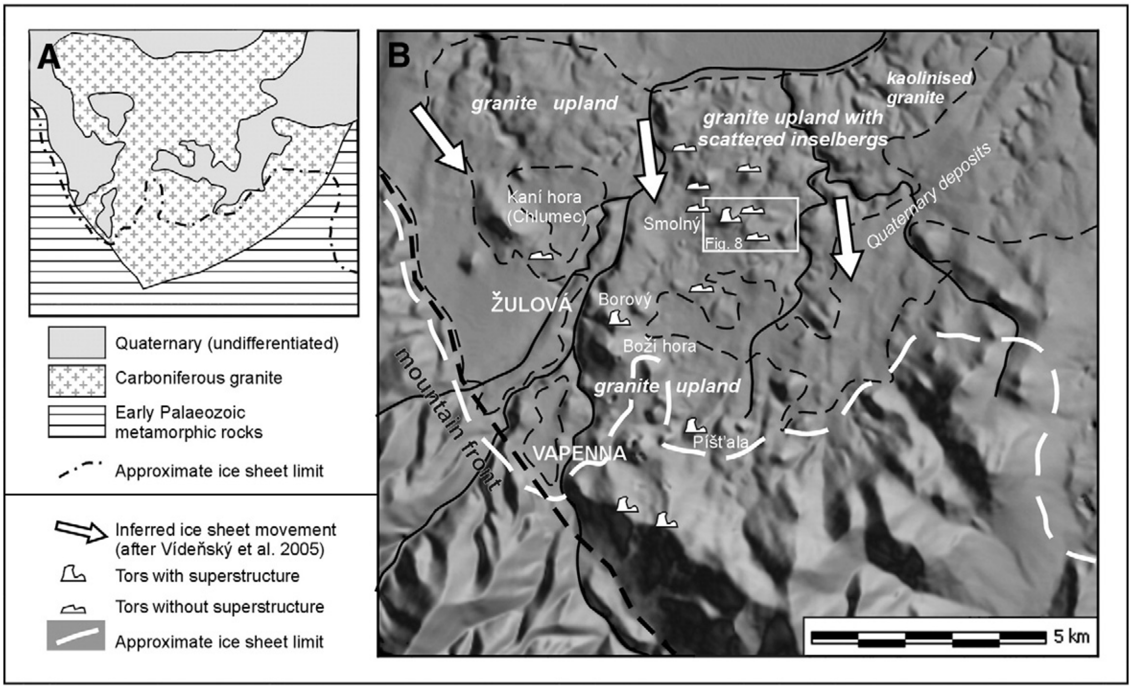

Fig. 6. Geology and geomorphology of the Žulova massif. (A) Geology. (B) Geomorphology.

direction in drainage lines and parallelism of valleys and ridges indicate strong structural controls on the granite relief. Sparse stream sections expose kaolinitic weathering beneath valley floors (Fig. 8), and this together with the 20–30 m depths of headwater valleys suggests that the granite hills are separated by deep weathering zones along fractures.

4.2.2. Interpretation

During the MIS 12, ice extended south of Žulova to terminate against the rising ground of the Sudetes along the southern margin of the massif (Sikorová et al., 2006). Glacial deposits have been recorded at up to 465 m asl east of Vapenná (Štěpančíková and Rowberry, 2008). This implies ice thickness of <100 m over dome

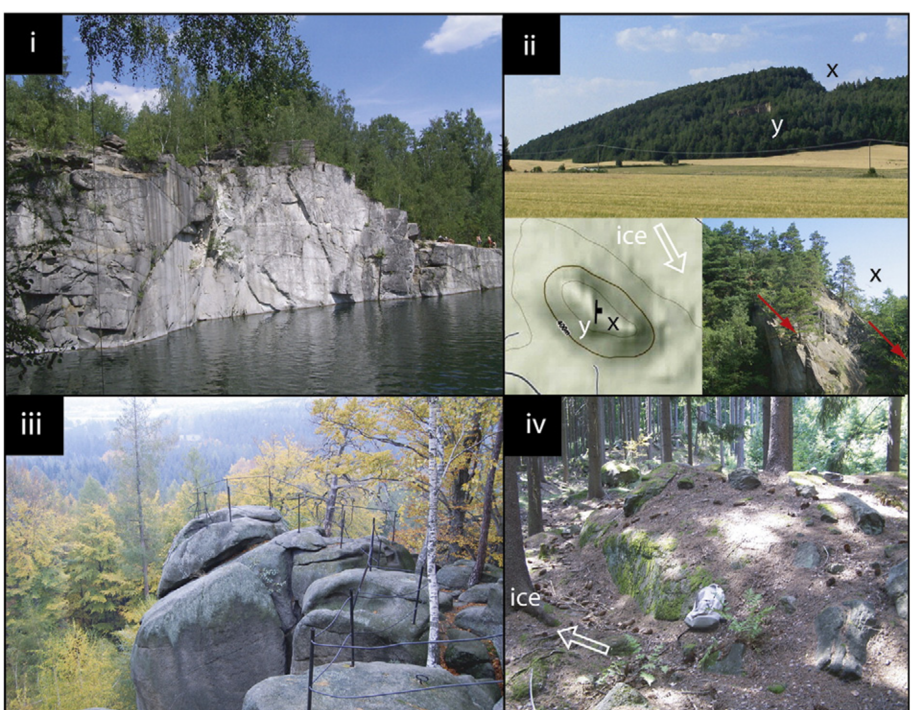

Fig. 7. Landforms of the Žulova massif. (i) NW margin of the Žulova Hilly Land. The hill flanks have a thin cover of till but otherwise this massive low dome has been stripped of regolith. (ii) The asymmetric dome, Kani hora. Although the hill is aligned with the direction of Elsterian ice flow, exposures on slope X and in old quarry Y show that the asymmetry is structurally controlled by sheet joints (red arrows). (iii) Summit of Smolny, with cliff edge and weathering pits over 1 m deep. (iv) Base of Smolny, with tor stumps showing incipient roche moutonnée-like forms.

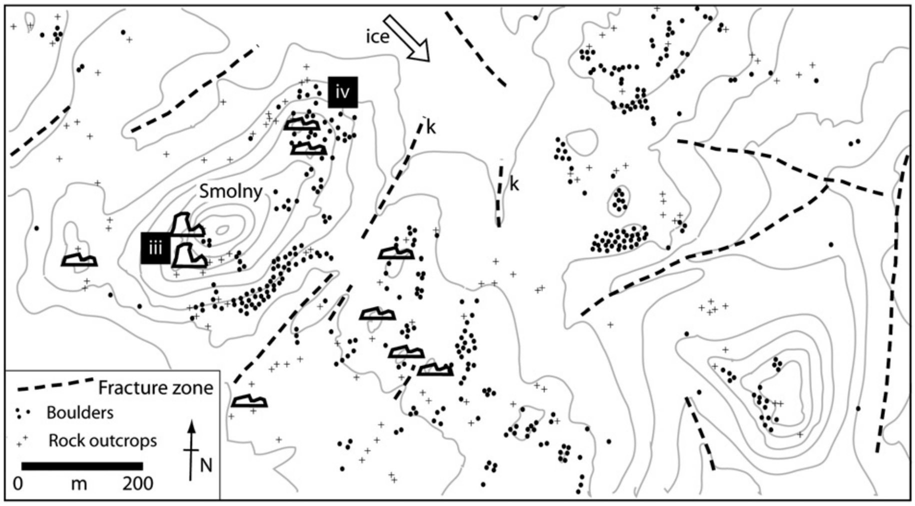

Fig. 8. Geomorphology of the area around Smolny. K = kaolinitic saprolite; tor symbols as in Fig. 3B. Locations iii and iv refer to Fig. 7.

summits and >250 m over low ground in the northern part of the Hilly Land, leaving Boži hora (525 m) as the only nunatak at the MIS 12 maximum (Fig. 2).

Demek (1976) recorded landforms in the Žulova Hilly Land that would fall within categories GEIs 1–5. He identified widespread stripping of regolith, interpreted features as roches moutonnées and attributed the asymmetry of hills, including Smolný vrch, to advanced glacial erosion. Later investigators, however, have regarded the hills, including low whalebacks, as structural landforms lacking in glacial elements (Ivan, 1983; Vídeňský et al., 2007).

In the hummocky terrain east and north of Smolný vrch, the evidence for glacial erosion is equivocal at best. Firstly, Pleistocene uplift of the Hilly Land has induced headward erosion of stream systems (Štěpančíková et al., 2008) and may have led to stripping of grus from slopes and the release of boulders, so there would be no need for glacial erosion to be invoked. Secondly, clearly developed roches moutonnées, with abraded slabs on the stoss side and well-developed lee-side cliffs, have not been observed in this study. Instead, low, parallel ridges ~5 m high, separated by shallow linear depressions, occur widely (Fig. 8). These ridges are locally bounded by rock steps, but ridge and step orientation is clearly controlled by joints and lacks any consistent relationship with the northwesterly movement of MIS 12 ice. The ridge tops often appear to have been cleared of boulders to reveal jointed rock surfaces, and boulder distribution does offer some support for the preferential transport of blocks by ice on to lee-side slopes (Fig. 8). The ridges occur close to small tors, however; and glacial stripping, if it has occurred, was limited and highly selective.

Firm evidence for significant glacial erosion on the domes is lacking. Massive low domes in the northern part of the massif lack regolith (Fig. 7) and may have been stripped by glacial erosion. The summits of large domes however retain delicate rock pillars, tors and micro-weathering features that predate glaciation (Demek, 1964; Štěpančíková and Rowberry, 2008). Dome flanks retain fragile tower tors and loose tor blocks (Fig. 7). The survival of tors and tor stumps at the base of the Smolný vrch and Kani hora domes cannot be reconciled with glacial modification of the adjacent hill summits. Dome asymmetry cannot be solely a product of glacial erosion because numerous quarry sections show that slope form is intimately controlled by sheet jointing (Fig. 7). The hills also lack streamlining in relation to the flow of ice from the NW. We conclude that dome elongation links to valley orientation along sets of NW–SE and NE–SW trending joints (Štěpančíková et al., 2008).

Again, the only likely GEIs in the Žulova Hilly Land are those for the earliest stages of glacial modification, with removal and reworking of regolith and blocks and minor modification of tors and low ridges. The domes show no clear evidence of glacial modification, aside from the removal of tor blocks.

4.3. Strzelin Hills

4.3.1. Evidence

The Strzelin Hills form a N–S trending ridge ~18 km long (Fig. 9). Second-order ridges, typically a few kilometres long, are aligned perpendicular to the main ridge. The overall relief is low. The highest elevation within the main ridge, Gromnik, is 393 m asl; whereas the valley floors to the east and west are located at c. 175 m asl. Slope gradients are low and rarely exceed 15°.

The Strzelin Hills are developed on metamorphic rocks of Proterozoic/Palaeozoic age, mainly orthogneiss, quartzite, and mica schist (Fig. 9A). These were intruded by a number of small granitoid bodies in the Carboniferous (347–330 Ma) (Oberc-Dziedzic et al., 1996). A larger outcrop area of granite occurs to the west and south of the town of Strzelin, having been extensively quarried; whereas an isolated stock builds the elevation of Gromnik.

The northernmost part of the Strzelin Hills displays subdued upland relief, reaching 200–250 m asl. Quarry exposures show that little regolith occurs under the tops and upper slopes of the granite elevations. Isolated pockets with 1–2 m thick disintegrated rock are occasionally exposed, but continuous layers of grus have not been observed.

In the southern area, tors become frequent above 340 m asl. On the northern slope of Gromnik, at 300–320 m asl, a few massive spur tors occur, with vertical walls 4–8 m high, lining a short gorge-like section of a headwater valley (Fig. 10). Angular boulders, apparently detached from the rock walls above, build a limited talus. The gneiss hill, Wyżna (370 m asl), has many low angular tors, 2–5 m high, separated from each other by scattered boulders. Few are upstanding on all sides; rather, the tors are asymmetric rock steps facing downslope. Of particular interest is one large glacially transported block, 10 m long (Fig. 10). No weathering microforms have been found on any of these tors. To the north of Gromnik, asymmetric quartzite cuestas project from the main ridge. Northwest-facing dip slopes are at 5–10°, considerably lower angles than the 20–25° of SE-facing scarps.

4.3.2. Interpretation

The least elevated, northern parts of the Strzelin Hills were fully exposed to ice sheet advances from the NW in MIS 12 and the NE in MIS 8 (Badura et al., 1998). The field evidence appears consistent with this situation. Neither tors nor deep regolith have survived in the

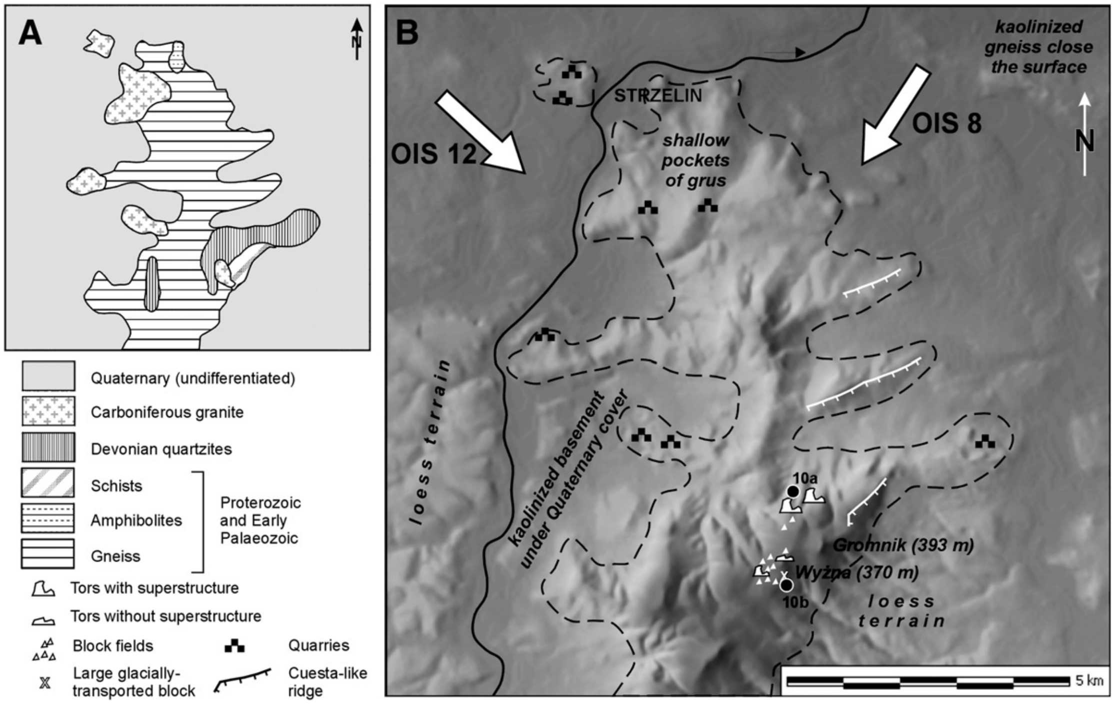

Fig. 9. Geology and geomorphology of the Strzelin massif. (A) Geology. (B) Geomorphology.

exposed northern end, suggesting complete stripping. However, the absence of rock steps on low granite domes means that no modification of the overall form of the hills can be recognised. In the southern area, the efficacy of glacial erosion was less, as testified by the survival of tor remnants above 340 m asl, but evidently sufficient to detach and move individual tor blocks. Localized examples of hill and ridge asymmetry, apparently consistent with moulding by ice advancing from the N and NW, are more likely from structural control. Evidence of hill asymmetry and streamlining as a result of glacial erosion remains elusive.

4.4. Strzegom Hills

4.4.1. Evidence

The Strzegom Hills form a cluster of subdued inselbergs, elongated NW–SE, in the NW part of the Sudetic Foreland (Migoń, 1997) (Fig. 1). Summits reach 275–351 m asl, and the height above the gently rolling footslopes varies between 35 and 110 m. Slope gradients typically are 10–15°. Kaolinitic saprolites, as much as 50 m thick, of likely Palaeogene to early Miocene age (Kural, 1979) occur under a cover of Neogene and Pleistocene deposits in topographic depressions between the inselbergs (Fig. 11).

The biotite granite of the Strzegom Hills has a well-developed, orthogonal fracture pattern. The dominant joint set trends to 140 ± 10°, and the secondary set to 50–60° (Podstolski, 1970); but it is the subhorizontal jointing that is visually most prominent (Fig. 12). Individual sheets of granite are typically 1–1.5 m thick, retaining this thickness over long distances. In very near-surface positions, further sub-horizontal discontinuities exist within the larger sheets, cutting the rock into thin slices <0.5 m thick.

Natural rock outcrops are rare and occur in two settings: on low gradient hilltop surfaces and on steeper slope segments, near the upper slope convexity. The former are no more than low slabs of relatively massive granite, rising above the grass-covered surfaces by 1–2 m at most. The latter are low (<5 m), angular cliffs, usually facing NW or NE. Exposed slabs appear heavily weathered, with very rough surfaces, projecting quartz crystals, and open vertical joints. Residual boulders are absent. A few low tors survive only on the lee slope of the small hill Jaroszowskie Wzgórza (Fig. 11). Here, the existence of these tors prior to at least MIS 8 glaciation is shown by the loss of superstructure to glacial erosion and by the recovery of a quartzite erratic from a löess-filled open joint. Pockets of disintegrated granite 4–5 m deep amidst more massive bedrock projections can be seen in a few quarries, but in exposed settings on hill slopes only a thin (0–1.5 m) mantle of grus is present, and the granite shows a generally planar weathering front (Fig. 12).

4.4.2. Interpretation

The altitude and geographical position of the Strzegom Hills require overriding by ice sheets. In the vicinity of the inselbergs, three separate till horizons have been documented and correlated with two ice advances during MIS 12 and one ice advance during MIS 8 (Krzyszkowski and Czech, 1995). The former are inferred to have been from the NW to N sector, whereas the latter would have been from N to NE.

The scarcity of tors and the very subdued nature of the remaining stumps may suggest glacial erosion and considerable remoulding of the preglacial surface. However, the importance of structural control cannot be understated. Quarry exposures show that dome slopes are adjusted to the inclination of sheet joints. Moreover, the fractured granite shows few signs of differential weathering and corestone isolation (Fig. 12). Hence, limited scope exists for tor emergence or boulder production, and the situation was probably similar prior to glaciation. Furthermore, the occurrence of rock cliffs in the stoss-side positions is difficult to reconcile with vigorous glacial erosion and is more consistent with limited truncation of hill spurs. On the lee side, a few detached slabs may be due to ice movement, but the lack of evidence for lee-side quarrying supports the view that the erosional capacity of the ice was limited.

The Strzegom hills show a NW–SE orientation. Whilst this is parallel to the direction of MIS 12 ice flow, it is unlikely however to be a product of glacial streamlining as adjacent hollows show the development of tens of metres of kaolinitic saprolites (Fig. 11A). Instead, the lineation appears to be primarily a reflection of basement

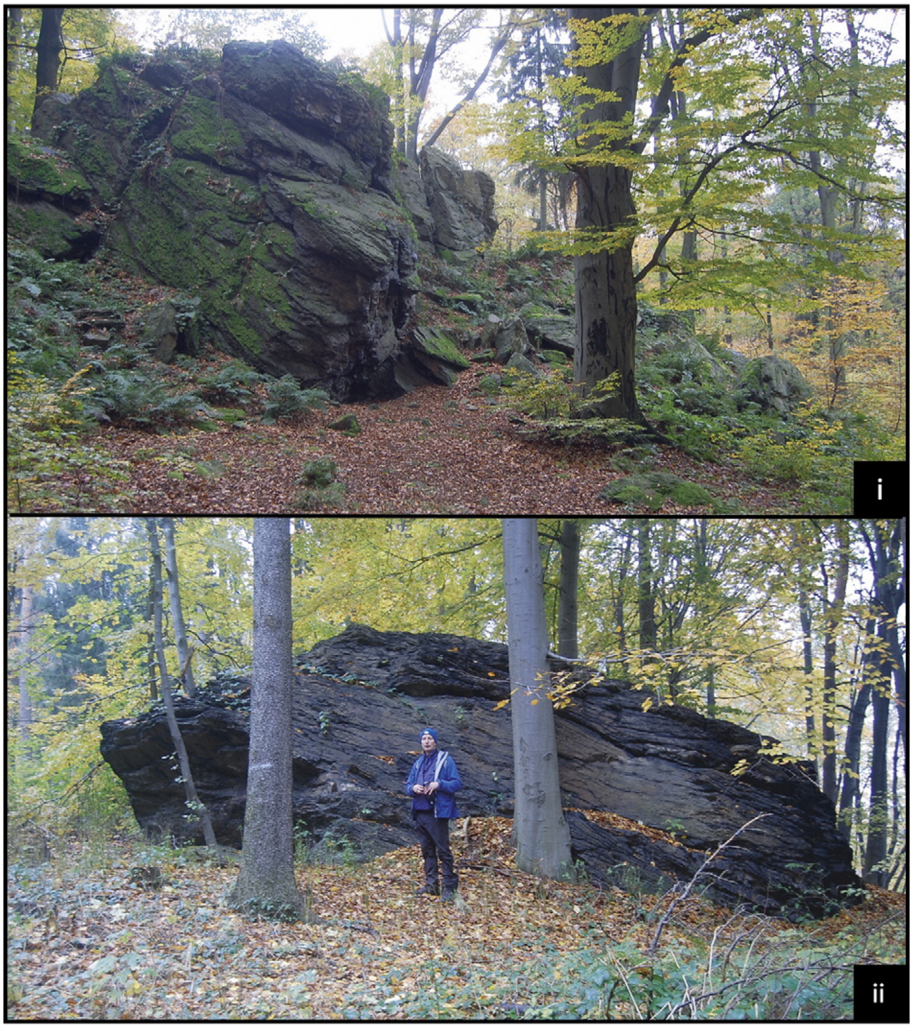

Fig. 10. Landforms of the Strzelin massif. (i) Tor from the valley head north of Gromnik; (ii) Glacially-transported tor block on the hill Wyżna, resting on block-rich slope deposits. Mineralogy of the gneiss matches that of tors found 180 m to the N.

faulting and joint orientation. Support for this interpretation comes from the cross-sectional form of the hills. Although northern slopes may be steeper, the many quarry sections show a parallelism with dome structure and a lack of rock steps or lee side cliffs (Fig. 12). The lack of deep glacial erosion on these domes is emphasised by the occasional occurrence of deep saprolite in pockets or broader zones several tens of metres across in quarry sections.

In the Strzegom-Sobótka massif, ice thickness probably reached 500 m in the MIS 12. Stripping of regolith and possibly tors from dome surfaces has been efficient but, again, there is no clear sign that glacial erosion of these massive domes has been capable of significant lee-side quarrying or streamlining.

5. Discussion

5.1. Impact of glacial erosion in the Sudetes and its Foreland

Models of glacier dynamics predict that the submarginal zone is one dominated by glacial deposition and that erosion is limited (Bennett and Glasser, 1996; Kleman et al., 2008). Few studies exist, however, of glacial erosion in resistant bedrock lowlands that lay beneath the margins of former ice sheets. Whilst some submarginal zones, such as SW Wales (Jansson and Glasser, 2008) and NE Scotland (Hall and Sugden, 1987) display few landforms of glacial erosion, others, including parts of New England (Jahns, 1943) and of Minnesota (Colgan et al., 2002) in the USA, display extensive rock surfaces that show striae, roches moutonnées, and other GEIs attributable to significant glacial erosion.

The Sudetic Foreland belongs firmly in the former category with few well developed glacial landforms. In the four study areas, we have identified only GEIs 1 and 2, features related to the stripping of regolith and the modification of small upstanding bedrock forms. GEIs 3 and 4 may be present locally, but we have been unable to document unequivocal examples of lee-side cliffs and roches moutonnées. GEIs 5 and 6 relate to the reshaping of large hills and none have been recognised in the Foreland. Further evidence of limited glacial erosion is provided by the survival of delicate pre-glacial features. On low ground, these include unconsolidated materials, such as Palaeogene kaolins, Neogene sediments, and granite grus that locally reach thicknesses of several tens of metres. On hills, these include small tors, blockfields, and weathering microforms. The inception of glacial erosion in the Sudetes has involved little more than partial and localised exposure of the basal surface of pre-MIS 12 weathering.

Within each study area the preservation of delicate pre-glacial landforms generally increases with altitude, a link that has been recognised in many other glaciated areas (e.g. Sugden et al., 2005). There is also a clear relationship between the impact of glacial erosion and distance from the ice margin. Close to the ice margin at Jelenia Gora and Žulova, numerous small tors have survived erosion beneath

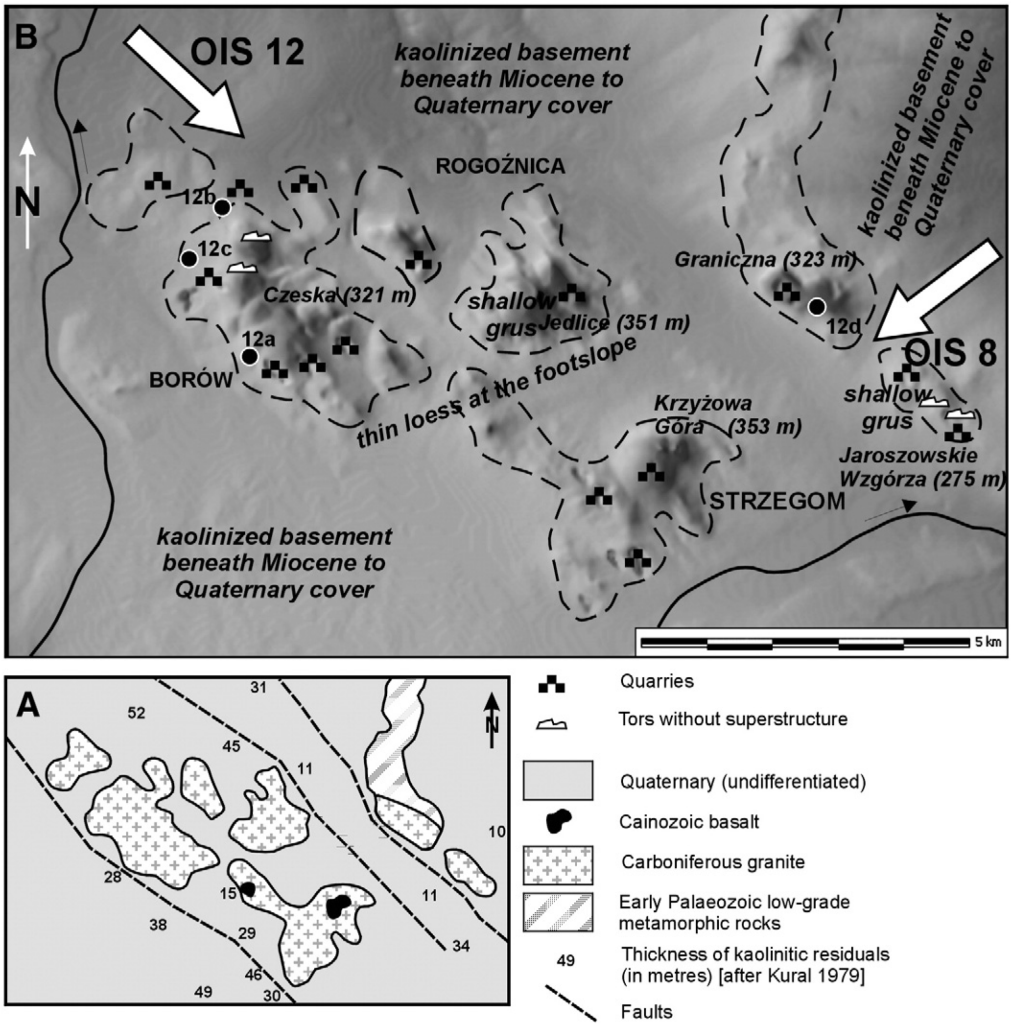

Fig. 11. Geology and geomorphology of the Strzegom Hills. (A) Geology. (B) Geomorphology.

ice thicknesses of >80–150 m; whereas around Strzegom and Strzelin, where ice thickness may have reached 500 m, these features are rare and confined to lee locations. Moreover, the total duration of ice cover was brief, amounting to <5 ka near the MIS 12 maximum limits and probably not exceeding 20 ka for the two or three glacial phases in the Foreland. Glacial erosion on the southern margin of the Scandinavian Ice Sheet was least where ice cover was thin and of shortest duration, a situation similar to that for the Laurentide ice sheet margin (Colgan et al., 2002; Stewart, 2007).

In contrast to the Sudetes, the granite landscapes that lay beneath the margins of successive Laurentide ice sheets in the NE USA have been largely swept clean of regolith (Goldthwait and Kruger, 1938) to leave resistant bedrock hills that show widespread evidence of glacial abrasion and quarrying. These areas lay beneath 500–1000 m of ice during not only the last but also earlier glacial maxima of the Pleistocene (Dyke et al., 2002; Marshall et al., 2002). The relative efficiency of glacial scouring in these landscapes, with extensive development of GEIs 4–6, can be attributed to greater ice thicknesses, higher flow velocities, and a much longer duration of ice cover.

5.2. Origins of glacial landforms

At the current early stage of glacial erosion, the Sudeten Foreland remains essentially an inherited pre-glacial terrain. This allows the first stages in the development of a number of important landforms of glacial erosion to be identified.

5.2.1. Tor modification by glacial erosion

Models of the progressive glacial erosion of tors have been developed in recent years (Rapp, 1996; André, 2004; Hall and Phillips, 2006). Application of the details of these schemes has not been possible in the present study because of the range of granite types and structures, the generally low density of weathering microforms and the dense forest cover. Nonetheless, the basic progression from castellated tor forms to subdued features and stumps is confirmed. Moreover, the survival of tors beneath ice cover is demonstrated unequivocally. The few tors that survive around Strzelin and Strzegom have experienced up to 20 ka of cover by hundreds of metres of glacier ice. Similar remarks apply to the massive granite tors of castle koppie type that exist at 300–400 m asl in the Lausitz area of NE Germany (Migoń and Paczos, 2007), ~25 km inside the Elsterian limit.

The evidence of tor modification, transport of tor blocks, and the discovery of erratic clasts all require that numerous tors existed with close to present morphologies before Elsterian glaciation at ~440 ka. This is consistent with cosmogenic exposure ages of 10^5 yr obtained for tor summits in other glaciated areas in Europe (Stroeven et al., 2002; Ballantyne et al., 2006, Ballantyne and Hall, 2008, Darmody et al., 2008, Phillips et al., 2006) and in North America (Briner et al., 2003). Some small tors may have emerged, however, since MIS 12

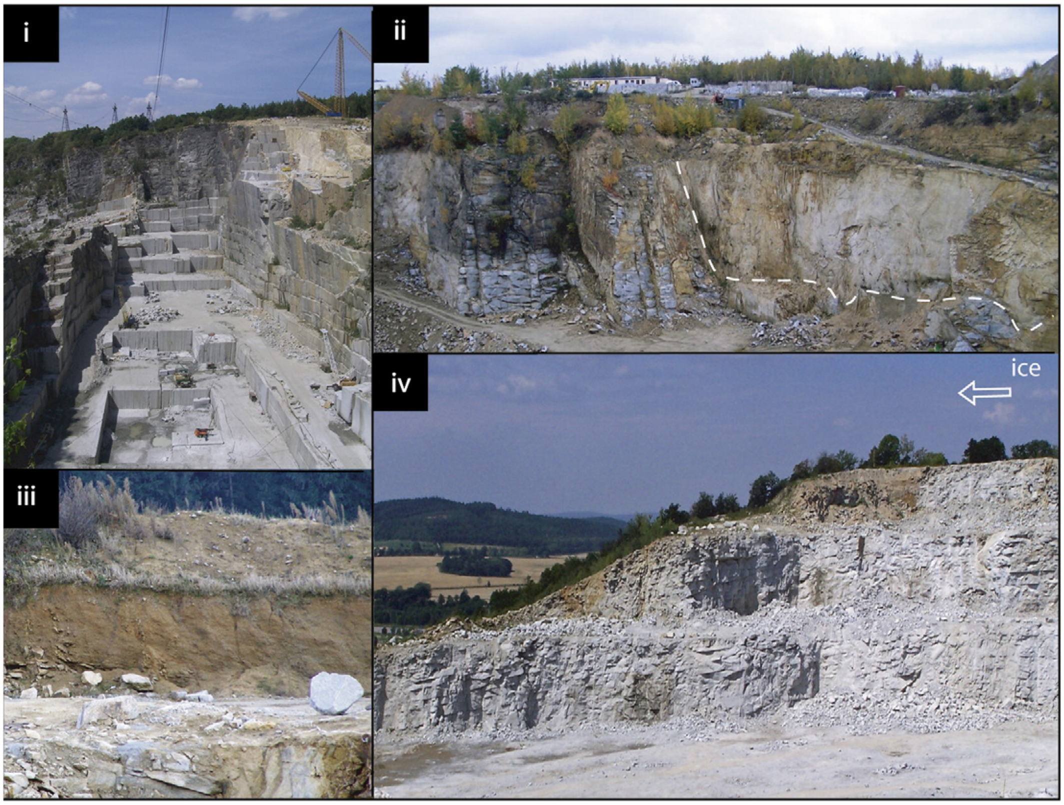

Fig. 12. Landforms of the Strzegom Hills. (i) Borów: quarry in flank of massive dome with sheet structure. (ii) Czernica: deep pocket of kaolinised granite. (iii) Gniewkow: thin (<2 m) granite grus at base of dome. (iv) Graniczna Hill (323 m): quarry section in lee slope showing absence of rock steps.

glaciation. For example, tors along the Bóbr gorge in the Jelenia Góra study area may have emerged through stripping of grus from slopes after valley deepening by meltwater erosion.

5.2.2. Roche moutonnée formation

Well-defined roches moutonnées have not been recognised in the study area, and development of streamlined glacier bedforms would appear to require a combination of greater ice thickness and longer glaciation than has been experienced in the Sudeten Foreland. Nonetheless glacial erosion of tors and other rock knobs has produced precursor forms. The inception of glacial erosion has involved entrainment of blocks and regolith to reveal jointed-bounded granite slabs. These bedrock masses have been further exposed by the partial removal of grus developed in surrounding zones of more closely jointed granite. Such incipient features however lack the abraded tops and lee-side cliffs typical of roches moutonnées. Similar forms have been identified in other areas of limited glacial erosion, including the Scandes of northern Sweden (André, 2001) and the Cairngorms, Scotland (Hall and Phillips, 2006). These findings provide support for long-standing views that one starting point for the formation of roches moutonnées is the stripping of the pre-glacial surface of weathering (Lindström, 1988; Olvmo and Johannson, 2002).

5.2.3. Inselberg modification by glacial erosion

The domed inselbergs of the Sudetes and its Foreland developed well before the Pleistocene glaciations, and are products of a long evolution, involving Miocene or older kaolinitic weathering, subsequent subsidence, and partial burial and renewed deep weathering and stripping under temperate environments in the Late Neogene. All the current elements of dome morphology, including shape, elongation, and slope form are inherited pre-glacial features that reflect strong lithological and structural controls. Quarry exposures show that granite domes are developed in compartments of relatively massive granite (Fig. 12), and that the angles of flanking slopes are closely defined by dipping sheet joints (Fig. 7); features that are also characteristic of granite domes in areas outside the limits of Pleistocene glaciation (Thomas, 1978; Römer, 2005). On the Strzegom-Sobótka and Strzelin massifs, tors and blockfields, where these existed, have been removed by glacial erosion from most summits, but dome surfaces remain parallel to closely-spaced sheet joints, and pockets of grus survive. Dome flanks have not been steepened, and occasional cliffs or steep slopes relate to steeply dipping sheet structures. Significantly, exposures in lee slope locations on hills do not show evidence of glacial quarrying and the production of rock steps. The onset of glacial erosion has failed to significantly modify the Sudetic domes.

In many studies of the glacial erosion of large granite hills, the depth of glacial modification has been estimated based on the twin assumptions that the pre-glacial hill shape was a simple dome and that the spacing and curvature of sheet joints matched its original surface (Jahns, 1943; Sugden et al., 1992; Glasser, 2002). The evidence from this study strongly supports those assumptions and confirms that the first stages of glacial modification of domes involve the stripping of loose material. The massive nature of the granite seen in quarries on dome flanks in the Sudetic Foreland is matched by experience in the NE USA (Jahns, 1943) and indicates that granite hills can be highly resistant to glacial erosion. This is consistent with the view that granite hills are often persistent, little-modified features in glaciated landscapes (Soen, 1965; Ebert and Hätterstrand, 2010). Moreover, questions are raised over the depth of glacial erosion needed to create a glacial landscape of asymmetric or streamlined hills from a pre-glacial hilly terrain (Olvmo and Johannson, 2002), especially where, as in parts of New England, kaolins survive at depth between glacially modified hills (Kerr, 1930: Kaye, 1967) and sheet joint patterns on granite hills imply limited erosion (Jahns, 1943).

The presence in the Sudetic Foreland of blockfields, block slopes and boulder-rich solifluction lobes above Elsterian and Saalian trimlines on granite, gabbro and basalt hill summits contrasts with the weak development of these features on glacially-stripped surfaces at lower elevations (Traczyk and Migoń, 2003; Migoń, et al. 2002; Żurawek and Migoń, 1999). Regeneration of these periglacial features on the massive, water-shedding rock surfaces of inselberg summits apparently has been very limited since 430 and 240 ka. The Sudetes and its Foreland, together with other areas of the world where the Pleistocene Glacial Maximum lay outside the Last Glacial Maximum ice limits, have great potential for further study of the rates of development of regolith, and saprolite and minor landforms over timescales of 10^5 ka under non-glacial conditions but such work probably requires detailed and comprehensive cosmogenic dating programmes.

6. Conclusions

Glacial erosion beneath the largely warm-based margin of the Scandinavian Ice Sheet in the MIS 12 and MIS 8 has had only limited impact on the terrain of the Sudetes and its Foreland. On hills, regolith was stripped by moving ice, tors were demolished, and blocks were entrained. Cold-based ice patches existed on the summits of large inselbergs, allowing the preservation of weathering microforms, blockfields, and tors. Indicators of more advanced glacial erosion, such as lee-side cliffs and glacial streamlining, are absent, however, even from granite domes that lay beneath ~500 m of ice. The survival of tors beneath ice cover, and the removal of tor superstructure by glacial erosion are confirmed. The presence of glacially-modified tors in areas covered only by Elsterian ice implies that the tors existed before 440 ka. Moreover, the contrast between blockfield-covered slopes found above Elsterian and Saalian glacial trimlines on summits of granite, gabbro and basalt hills in the Sudetic Foreland and glacially-stripped surfaces at lower elevations implies only limited regeneration of blockfields since 430 and 240 ka. The first stages of roche moutonnée formation are recognised in the exhumation of tor stumps from grus but development of lee-side steps remains very limited. Granite domes retain a pre-glacial morphology determined by strong structural control over hill form. Glacial erosion increases with distance from the former ice margin but the weak development of roches moutonnées and of asymmetric and streamlined hills even within areas where the Scandinavian Ice Sheet was ~500 m thick and persisted for ~20 ka indicates that development of these features of advanced glacial erosion requires significantly thicker or more persistent ice cover.

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