Styles of soft-sediment deformation on top of a growing fold system in the Gosau Group at Muttekopf, Northern Calcareous Alps, Austria: Slumping versus tectonic deformation

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Abstract

Styles of soft-sediment deformation in a syngrowth coarse-grained sediment gravity flow system are described from outcrops of the Upper Gosau Subgroup of the Northern Calcareous Alps. The deepwater sediments were deposited along the flanks and top of growing folds. Beside load and fluid escape structures, which are commonly found in sediment gravity flow systems, soft-sediment deformation is mainly related to fluidization of coarse-grained beds and downslope gliding of meter-thick sediment packages. Common structures affecting single or several beds are: hydroplastic folding of fine-grained beds in a coarse-grained fluidized matrix, symmetric or asymmetric mullions at the base of coarse-grained beds, and folding of heterolithic units into coarse-grained beds. Folds have varying geometries and styles, and fold axes are scattered. Soft-sediment folds also affect thick sediment packages. These folds are uniformly verging asymmetric metric to dekametric folds, formed at different degrees of lithification. Rheology of sandstones in such folds ranges from ductile to semi-brittle and brittle. Fold axes tightly cluster about one direction. The first group of deformational structures is attributed to slump-related deformation, whereas the second group represents folds related to tectonic deformation.

A dislocation model and a shear zone model are used to interpret orientation of folds related to slump deformation. This considers processes taking place on the scale of the whole slump sheet and within the shear zone at the base of gliding units. Both models predict rotation of fold axes toward the transport direction of the slump. Field data support that fold axes scatter either about the downslope transport direction or about the strike of the slope.

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1. Introduction

Dating of deformation has been an important issue in structural geology in the last few years, and important progress has been made in dating of deformation in metamorphic rocks by dating single grains that grew synkinematically (e.g. Müller et al., 2000). Dating of deformation in shallower levels of the crust is not possible by geochronologic methods, as temperatures are too low for recrystallisation. If sedimentation is contemporaneous with deformation, the maximum and minimum age of fault-related deformation can be defined: the maximum age of deformation is given by the age of the sediment in contact with a fault, the minimum age by the age of the sediment overlapping a fault. Direct dating of deformation is possible in sedimentary successions if larger scale geometries such as progressive unconformities are present (Riba, 1976; Anadon et al., 1986; Medwedeff, 1989; Ford et al., 1997; Suppé et al., 1997). In these cases, the age of
the sediment, determined by paleontological methods gives the age of deformation. Sedimentary successions deposited on growing folds will be subject to (1) loading and dewatering intrinsic to deposition, (2) gravity-driven mass wasting processes such as slumping due to progressive rotation of fold limbs in addition to (3) tectonic deformation. Because both tectonic deformation and slumping affect unlithified or poorly lithified sediment, similar small to medium scale structures will form. This paper presents field observations on soft-sediment deformation from deep water sediment gravity flow deposits and discusses their origin, with the goal to define criteria for the distinction between slump-related deformation and tectonic deformation in the field. The unusual wide range of grain sizes in a wide variety of gravity flow deposits from megabreccias to marls in a growth strata setting make the Gosau Group at Muttekopf an excellent natural laboratory for characterizing and testing the ability to discriminate gravity versus tectonically driven processes and products.

1.1. Geological setting

This paper deals with Upper Cretaceous sediments of the Gosau Group at Muttekopf in the western part of the Northern Calcareous Alps of the European Eastern Alps. The Gosau Group is an Upper Cretaceous, synorogenic carbonatic–siliciclastic sedimentary succession, which unconformably overlies deformed Triassic to Jurassic rocks (Wagreich and Faupl, 1994; Sanders et al., 1997). Generally, deposition started in a terrestrial environment, which subsided to neritic conditions (Lower Gosau Subgroup). After a pronounced subsidence event deep marine conditions prevailed (Wagreich, 1993), and the Upper Gosau Subgroup was deposited. The relationship between the contracting orogenic wedge and the coeval major subsidence is not well understood at present, and different models have been put forward (e.g. Wagreich, 1993; Froitzheim et al., 1997).

The younger, deep marine part of the sedimentary succession (Upper Gosau Subgroup; Wagreich and Faupl, 1994) was deposited during transport of the thin-skinned nappes of the Northern Calcareous Alps over tectonically deeper units (Fig. 1). In the Muttekopf area, internal deformation of the moving nappe led to the formation of fault-propagation folds in the subsurface of the Gosau sediments and hence to formation of several (progressive) unconformities within the sedimentary succession (Ortner, 2001; Figs. 1–3). The Upper Gosau Subgroup is divided into three sequences (Ortner, 1994, 2001; Sullivan et al., 2006; Fig. 2): All three sequences of the Upper Gosau Subgroup are dominated by vertically-stacked, upward-finishing, laterally continuous, unchannelized conglomerates and sandstones that display little to no lateral variation in facies. The boundary between Sequence 2 and 3 is the Rotkopf unconformity (Figs. 1–3). The boundary between Sequence 1 and Sequence 2 is the base of the 2nd fining-upward sequence, which is significantly below the most prominent unconformity in the area (Schlenkerkar unconformity) (Figs. 1, 2). In Fig. 2, Sequence 2 includes both offlapping stratal geometries, suggesting that the rate of structural growth was greater than the rate of deposition and progressive onlapping stratal geometries, suggesting that the rate of deposition was greater than the rate of structural growth. Due to limited preservation of Sequence 3, its stratal geometries are less clear. Sediment transport directions are either parallel to the regional fold axial trend and are southwest or northeast-directed, or perpendicular to the structural trend and north-or south-directed (Ortner, 1993, 1994). The direction of dip of the paleoslope will be reevaluated in the discussion (Section 3.2).

The style of tectonic deformation changes from west to east: in the western part of the outcrop, the Gosau sediments are preserved in the core of a km-scale syncline (Muttekopf syncline), which is interpreted to be the leading syncline of a fault-propagation fold formed above a deep-lying blind thrust (Fig. 2). Further east, the southern part of the Gosau outcrop is characterized by dekametric folds, which formed above minor blind thrusts which are partly exposed (Fig. 3). Structural style changes across NW-striking tear faults (Fig. 1). Fold growth of the main fault-propagation fold was syndepositional and is documented by a progressive unconformity in the southern limb of the leading Muttekopf syncline (Schlenkerkar unconformity of Figs. 1, 2). Growth of the fold north of the Gosau outcrop, which led to southward tilting of the northern limb of the Muttekopf syncline is mainly post-depositional (Ortner, 2001), only minor stratal convergence can be seen in Fig. 3. Fold growth in the zone of complex folding at the southeastern margin of the Gosau outcrop east of Rotkopf (Fig. 1) was also syndepositional and led to rotational onlap at the Alpjoch unconformity (Figs. 1, 3).

1.2. Introduction into soft-sediment deformation

Soft-sediment deformation structures are common to most sediment gravity flow systems. Gravity flows are mixtures of water and sediment and are often rapidly deposited due to flow collapse. This rapid deposition traps interstitial fluid in the sediment and results in high, unstable pore pressures. Most soft-sediment deformation is related to fluid-escape due the squeezing out of excess water from the pores of a sediment during
Fig. 1. Geological sketch of the study area. Inset a: Position of the Muttekopf Gosau outcrop in the Northern Calcareous Alps and Austria. Inset b: Nappe units of the Eastern Alps in the wider vicinity of the investigated area. 1: Schlenkerkar unconformity, 2: Rotkopf unconformity, 3: Alpjoch unconformity. Modified from Ortner (2001).
compaction. The style of soft-sediment deformation structures changes with progressive burial, going along with a change of the main controlling factors.

In unconsolidated sediments near sediment–water interface pore pressures can equilibrate by either gradual or rapid fluid escape. Rapid dewatering in many cases is...
Fig. 3. Panoramic view of the eastern part of the Gosau outcrop from the Platteinwiesen in the west. Note the zone of complex folding in the southwestern part of the Muttekopf outcrop, related to minor thrusts reaching into the sedimentary succession.
triggered by rapid deposition of a dense sediment (e.g. sand) over a less dense, water-rich sediment (e.g. mud), creating a reversed density stratification. Under these conditions pore water pressure rises above overburden pressure, the underlying low-density sediment is fluidized (see below) and intrudes the overlying unconsolidated sediment. Structures created by such a process are load casts and flame structures (ball-and-pillows and convolute bedding; Anketell et al., 1970; Lowe, 1975; Visher and Cunningham, 1987). Dewatering in the shallow subsurface triggered by seismic shocks is another reason for the formation of dewatering pipes (Owen, 1995). If sediment dewatered gradually near the sediment–water interface under its own weight, pillars and dish structures form (Lowe and LoPiccolo, 1974; Lowe, 1975).

A more general description of soft-sediment deformation, which can also be applied to structures related to subsurface sediment mobilization was given by Knipe (1986). He described deformation type as a function of three major parameters: deformation ratio, degree of lithification and fluid pressure. Three types of deformation were defined: independent particulate flow, diffusion mass transfer and cataclasis.

(1) Independent particulate flow describes the behaviour of non-cohesive or poorly lithified sediments. Depending on the ratio of excess pore fluid pressure to cohesive strength due to grain weight, three deformational processes are defined: hydroplastic deformation, liquefaction and fluidization.

(A) Hydroplastic deformation occurs when fluid pressure is lower than grain weight and preserves but modifies primary sedimentary structures such as bedding. Many structures created during hydroplastic deformation resemble plastic deformation in metamorphic rocks, and can be described and interpreted using the terminology of structural geology.

(B) When fluid pressure is equal to grain weight, the sediment liquefied and is subject to laminar flow. Primary sedimentary features such as bedding are destroyed.

(C) When fluid pressure is higher than grain weight, or fluid velocity is high enough to entrain grains in the fluid, the sediment fluidizes and is subject to turbulent flow. All primary sedimentary structures are destroyed. Fluid pressure increases when low-permeable sediment seals a high-permeable sediment below and prevents fluid escape. Fluid pressure then approaches lithostatic pressure (Jolly and Lonergan, 2002). If fluid pressure exceeds both tensile strength of the seal and the principal horizontal stress, which is the minimum principal stress an undisturbed system, a clastic dyke forms. In cases when the principal vertical stress is smaller than the tensile strength of the seal, and fluid pressure is intermediate, a sill forms.

(2) Diffusion mass transfer and (3) cataclasis were also observed in partly lithified rocks, but do not play an important role in this study.

2. Description of soft-sediment deformation structures

The following section describes centimetric-to hectometric deformation features observed in the field and provides a short interpretation. Deformation structures that formed near the sediment–water interface are described first, followed by structures that formed with greater sediment overload. Common soft-sediment deformation structures such as load casts, which are abundant in the study area, are not described.

2.1. Structures affecting single or few beds

2.1.1. Loading

On top of deformed sand–mud couplets, elongate troughs filled by coarse-grained sediment were observed (Fig. 4). When tracing the sand-beds of the sand–mud couplets, it appears that the thin sandstone beds slightly diverge when moving from below a trough to the crest of a rise (Fig. 4). Moving upsection in the thin-bedded unit, amplitude of the waves increases from zero to a maximum of about a meter, and locally the tops of the rises have teepee-like geometry, or, in other words, the geometry of upward pointing cusps (Fig. 4a). The sediment fill of the troughs is horizontally laminated, and the laminae onlap the lateral margins of the trough (Fig. 4b). Toward the top, the coarse-grained unit grades to sand, and the top of the bed is flat (Fig. 4a). The long axis of the troughs is oriented almost perpendicular to the sediment transport direction indicated by flute casts in neighboring sandstone beds (see diagram in Fig. 4a).

The flat top and the sideward onlap of laminae of the coarse-grained bed suggest that this sedimentary units filled preexisting topography. Undeformed horizontal laminations in the bed preclude loading by the coarse-grained material, and suggest the troughs should have existed before deposition of the coarse-grained unit. On the other hand, the absence of a basal glide plane, above which folding of the sand–mud couplets could have taken place, and progressive upward development of the wavy folds by divergent and convergent bedding
indicates differential compaction and therefore loading. A possible interpretation of the features observed suggests initial differential loading by sand probably organized in sand waves, which caused the wavy surface observed. Higher current velocities during a second event enabled the current to remove the sand and deposit the gravel onto the existing topography.

2.1.2. Fluidized coarse-grained beds

A significant part of the sedimentary succession of the Upper Gosau Subgroup of the Muttekopf is built by meter to several meter-thick conglomerate beds. Usually, these beds exhibit little internal organization. Frequently, clasts of finer grained sediments (coarse sandstones) float in the unorganized conglomerate (Fig. 5a,b). Locally, conglomerate beds show fluid escape structures (flame structures) into overlying sandstones. Fig. 5a shows a conglomerate flame structure that has vented through a 10 cm sandstone bed to feed a bed-parallel injection (sill). Conglomerate intrusions may also have a diapiric form (Fig. 5d) with vents that crosscut sandstone beds and indicate fluidization. Overlying sandstone beds are only weakly deformed, and horizontal lamination preserved, indicating hydroplastic deformation of the sandstone. Where an initial continuous grading from conglomerates to sandstones and siltstones was present in a bed, only the central part of the bed was fluidized. The upper and lower boundaries of fluidization are not strictly parallel to bedding. In Fig. 5b, the boundary between the fluidized unit and overlying siltstones follows planes of a conjugate joint set, which are related to layer-parallel shortening, because the acute angle between conjugate joints contains the bedding plane orientation.

Two important conclusions can be drawn from these observations: (1) Fluidization of the conglomerate bed occurred after burial of the now fluidized bed below younger sediments. Increasing overburden pressure during ongoing sedimentation may have contributed to rise of pore fluid pressures needed for fluidization of conglomerates (compare Lowe, 1976). (2) As grain weight in the conglomerate beds was supported by moving pore fluid during fluidization, and the sandstone bed with its much smaller and lighter grains was not fluidized, the ratio between grain weight and pore fluid pressure is not the sole control on fluidization. Sandstones were probably cohesive due to a higher content of clay minerals or due to incipient cementation or both.

Fluidized coarse-grained beds can have varying amounts of incorporated fine-grained beds. The sand-or siltstone beds are disrupted and folded, but internal structures like grading and lamination are well preserved (Fig. 6a). The organization of the sandstone beds within the fluidized layer depends on the amount of coarse matrix and on the size of the components floating in the matrix relative to bed thickness:

(1) Where large rafts of fine-grained sediment beds are found, shingle-like stacking of sandstone beds is observed beside folding (Fig. 6, fluidized layer 1). The poles to bedding of sandstone beds are arranged in a girdle perpendicular to the sediment transport direction, and parallel to the downslope direction, in the case the folds within fluidized layer 1 facing south indicate the downslope direction. Many folds are isoclinal, and are refolded by open folds (Fig. 6). Fold axes (Fig. 6d) are dispersed about a WSW trend, and most are parallel
to bedding. Axial planes of folds are inclined with respect to bedding, and most of them strike WSW, parallel to the mean axial trend (Fig. 6b).

(2) Where fine-grained material is broken up to clasts much smaller in size than the thickness of the fluidized layer, such as the one shown in Fig. 6, fluidized layer 2, folds are near isoclinal, and axial planes are all near parallel to bedding (Fig. 6g). The fold axes scatter around the slope dip direction (Fig. 6f) and can be strongly curved (sheath folds).

(3) Where matrix dominates the fluidized bed, fold axes within the isolated clasts are randomly oriented.

In all cases, folded sandstone beds show parallel fold style, irrespective of the interlimb angle. The mechanism of folding was flexural folding as shown by the occurrence of mullions in the inner bend of isoclinal folds where more competent sandstone is in contact with mudstone, and minor normal faults at the outer bend (Fig. 6a). No similar style folds were observed in fluidized layers.

Incorporation of sandstone beds into fluidized conglomerates was probably a result of conglomerate intrusion into overlying and underlying sandstone beds. Coalescing of neighbouring conglomerate sills would isolate parts of sandstone layers and incorporate them into a fluidized bed. Intrusions across several sandstone layers or multiple intrusion events could produce sandstone-rich liquefied layers. Once incorporated, the sandstone beds could probably be further disrupted by intrusions across layers, as demonstrated by Fig. 5c, which shows part of a conglomerate dyke crosscutting a mudstone clast-within the conglomerate layer.

Fig. 6 shows the upslope end of a fluidized layer (right end of fluidized layer 1). There, a thrust plane cuts upsection and ends below fluidized layer 1. Immediately above the thrust tip within the fluidized layer, a fold in a package of sandstone layers faces approximately south, and is covered by a conglomerate rapidly wedging out laterally. Fluidization lasts only for very short time intervals (e.g. Lowe, 1976). In fluidized state, sediments should move downslope. Any mass transport on a slope has an upslope part, from which material is transported toward the downslope part, and the upslope part should be subject to vertical thinning and horizontal extension, whereas the downslope part should experience shortening.

Fig. 5. Evidence for fluidization of conglomerate beds. a: Flame structures at the interface between sandstone and conglomerate projecting into the sandstone. To the right conglomerate was injected upwards into sandstones forming a bed-parallel sill. Clasts of coarse sandstone give evidence of mixing in an originally continuously graded bed. Length of hammer head 20 cm. b: conglomerate dyke injected into mudstone, which is a large clast within the conglomerate. c: fluidization in a bed continuously graded from conglomerate to siltstone. Fluidization is restricted to the central part of the bed, where silt-and sandstone clasts float in a conglomerate matrix. The capping siltstone is cut by conjugate joints. Joint surfaces delimit the fluidized unit against the siltstone. d: conglomerate intrusion into coarse sandstone coupled with large dish structures within the sandstone. Scale is 7 cm.
Fig. 6. Key outcrop of two fluidized beds near Muttekopfhütte. Fluidized layer 1 formed at the tip of a minor thrust cutting upsection from right to left. At the right end of the outcrop, fluidized layer 1 is refolded by semi-brittle soft-sediment folds, which are depicted in detail by in Fig. 9a. a: isoclinal fold within fluidized layer 1. Grading and lamination in sandstone bed is well preserved. b: poles to bedding of deformed sandstone bed within fluidized layer 1 and calculated fold axis. c: axial planes of folds in deformed sandstones within fluidized layer 1 and calculated fold axis. d: trend and distribution of small-scale axes of folds in deformed sandstones at fluidized layer 1. e: sediment transport directions from flute casts and tool marks at the base of sandstones and conglomerates. f: axial trends and and distribution of small-scale folds in deformed siltstones within fluidized layer 2. g: deformed heterolithics and injected sandstones in fluidized layer 2. h: sketch of upramping of a unit during fluidization.
and thickening. The observation of folding and hence shortening at the upslope end of a fluidized layer is unusual. Fluidization might be a result of initial movement at the thrust by compressing the pore water in a layer sealed at the base and the top (compare joints of Fig. 5b). Pore water pressure would then reach the critical value for fluidization. Because the wedging conglomerate layer at the top is not folded, the fold at the upper end of the fluidized layer must have been created during movement of the thrust and during fluidization. In surficial slumps, such a relationship between downslope gliding of coherent units, which is described in Section 2.2.4, associated with fluidization of the lower part of the slump was observed by Loope et al. (2001). Normal faulting in the units on top of fluidized layer 1 might be a result of downslope gliding and associated stretching of the overlying sediment package during fluidization.

The thickness of fluidized layer 1 increases in the downslope direction (Fig. 6). Decrease of thickness in a fluidized layer of another outcrop is associated with a verging fold in the sandstone layer sealing the fluidized unit at the top, giving evidence of gliding during fluidization (Fig. 6h). Where the sandstone layer was doubled, the sandstone projecting into the fluidized unit was assimilated into the fluidized unit.

2.1.3. Hydroplastic deformation at the base of coarse-grained beds

Where a sandstone beds underlies a coarse-grained bed, the interface with fine-grained deposits such as sand–mud couplets is usually undeformed (Fig. 6). However, where fine-grained sediments are in direct contact with conglomerates, the contact is deformed, as is demonstrated by a conglomerate layer with a boudined sandstone bed at the base (Fig. 7a). Common deformational structures at such contacts are decimetric to metric elongate flame structures of silt–and mudstones into conglomerates. Flame structures can be isolated, with distances of several meters from one flame to the next, or closely spaced. In the latter case, elongate troughs are developed between individual flames. In many cases, these troughs and associated flames are symmetric and slightly irregular (Fig. 7a), however in other cases, both are highly asymmetric and consistently verge in one direction (Fig. 7c,e). Unlike in Fig. 5a, where flames show evidence for fluid escape from conglomerates into overlying sediments, the vergent flames developed at the base of conglomerate layers rather resemble folds (Fig. 7f). Centimetric alternations of silt–to mudstones are folded in open to isoclinal folds into the conglomerate.

Where flames are asymmetric, a glide horizon was observed within the silt–to mudstones, and the sediments above the glide horizon are deformed (Fig. 7c,e,f). Directly above the glide horizon, isoclinal folds with axial planes subparallel to bedding are abundant, and fold style in deformed sand–and siltstones is similar. The isoclinal folds are frequently boudinaged, and resemble intrafolial folds as observed in metamorphic rocks. “Flames” are developed, where shear zones branch off the glide plane parallel to bedding and reach into conglomerates. Folds associated to the latter shear zone are not boudinaged, only locally the steep limbs of verging folds are sheared.

For interpretation, the observed structures at the base of conglomerate beds must be split up into several subtypes. Boudinage of a sandstone at the base of a conglomerate indicates stretching of the sandstone and conglomerate beds and boudinage of the sandstone due to higher viscosity of the sandstone. Initial boudinage is probably a result of downslope gliding and stretching during initial fluidization of the conglomerate layer, as indicated by local conglomerate intrusions (Fig. 7c). Where the sandstone was removed due to boudinage, conglomerate came into contact with fine-grained sediments.

Loading of fine-grained sediments by the conglomerate could create load casts. Loading as a consequence of superimposing a dense plastic or fluid layer on a substratum of lower density should create balls-and-pillows separated by flames arranged in a polygonal pattern (Owen, 2003). If loading was contemporaneous with downslope translation, elongate ridges of low-density sediment projecting into high-density sediment parallel to transport direction would be expected (e.g. Anketell and Dzulynsky, 1968; Moretti et al., 2001). Dzulynsky (1966) suggested convection-like particle movements during transport of sand deposited by turbulent currents, to explain ridges parallel to transport direction found at the base of beds. For conglomerates and breccias, which were transported and deposited by debris flows or high-density turbidity currents, such structures were not described.

As the flames and balls do not seem to be related to sedimentary processes (see above) and are perpendicular to sediment transport direction (Fig. 7a), an alternative interpretation is needed. Folding of the interface between conglomerates and fine-grained sediments would be a possible explanation. A prerequisite for this interpretation is that both lithologies deform (hydro-)plastically, and hence it has to be assumed that the structures postdate liquefaction. The geometry of the folded interface can be described as cuspatelobate, with the cusps developed in the fine-grained material and the lobes in the conglomerate. In structural geology terms, such folded interfaces are called mullions (e.g. Urai et al., 2001). According to these authors, “mullions are described as
regular cuspate–lobate folds of an interface between two materials with a large competence contrast, the cusps pointing towards the more competent material. At the time of formation of the mullions, conglomerates must have been more competent than the silt- and mudstones below.

As the mullions are more or less symmetric, they record pure shear shortening parallel to bedding. Where asymmetric mullions are found (Fig. 7c,e), either formation of mullions by simple shear must be assumed, or initially formed symmetric mullions were overprinted by simple shear. Fig. 7f shows an example of an isolated “cusp” into a conglomerate bed. There, a shear zone branches off the basal glide plane of the conglomerate bed. Isoclinal folding with axial planes parallel to the basal shear zone and subsequent boudinage of isoclinal
folds can be seen as an effect of progressive simple shear. Initially, folds above the glide plane formed with SE-dipping axial planes, but were progressively rotated into parallelism with the shear zone and then extended.

The strain histories of individual conglomerate layers can be complex, as seen from the descriptions and interpretations above. Initially, conglomerates became stretched due to downslope gliding on top of a glide plane which remained active throughout early deformation of the bed. Then bedding parallel shortening became active leading to formation of symmetric or asymmetric mullions. The mechanism determining the type of mullion developed remained unclear at the moment. As can be seen in Fig. 7d, the position of the outcrop in the syncline is not the controlling factor, because different types of mullions are found in the same bed in one limb of the Alpjoch syncline.

2.1.4. Hydroplastic deformation within coarse-grained beds

The main difference between units deformed by fluidization and units affected by hydroplastic deformation is the absence of disruption of sandstone beds and the absence of pervasive mixing of conglomerates in the latter. In Fig. 8, the lateral margin of a graded breccia bed affected by slumping is depicted. Several isoclinal folds of a heterolithic unit consisting of sand–mud couplets originally underlying the breccia reach into the breccia bed. The axial planes of the folds are subparallel to bedding and the axes of the folds are strongly curved. Folds have sheath fold geometry (e.g. fold in the center of the Fig. 8, orientations of fold axis represented by squares in Fig. 8c). Anastomosing shear zones within the conglomerate bed originate at the hinge of isoclinal folds.

The isoclinal fold in the center of Fig. 8 roots in the folded zone at the right side (diagram Fig. 8d). Therefore the sequence of deformations resulting in the observed structure is probably as follows: Initially, an asymmetric mullion as described in Section 2.1.3 developed. When the hinge of the progressively amplifying verging fold reached the interface between coarse breccia and fine breccia, which acted as a rheologic interface, further fold growth was guided by the rheologic interface. Probably the mechanism of fold growth changed at the time the hinge zone reached the rheologic interface from folding by subhorizontal shortening and progressive tilting of the fold limbs to gliding of the unit above the rheologic interface and rolling the heterolithic beds attached at the base through the hinge zone of the isoclinal folds. Taking the axis and the facing direction of the conical fold in the heterolithic unit at the right side of Fig. 8 as an indication for direction of slumping, the slump moved to the southeast. Fold axes measured in the isoclinal folds within the slumped bed are dispersed, but scatter about the direction of slumping (Fig. 8c).

2.2. Structures affecting thick sediment packages

In contrast to the structures described up to this point, which affected single layers of the sediment succession, the folds which will be described in the following paragraphs reach over sediment stacks of 10 to 100 m thickness.

2.2.1. Metric folds

At the left side of Fig. 6, a train of folds deforms fluidized layer 1. Fig. 9 depicts one of the folds below fluidized layer 1 in detail. All folds in the outcrop are asymmetric and face N. The axial planes of these folds are parallel and inclined with 25° against bedding (Fig. 9a). Fold axes of the folds are well aligned and trend E–W (Fig. 9b). Axial planes of folds can be traced across 3–4 coarse-grained layers (Fig. 6, lower left). Sandstone beds are frequently faulted and stacked, and on the fault planes stretched calcite fibers indicate the movement sense (Fig. 9c). Some faults were refolded in the duplex in the steep limb of the fold. Locally individual sandstone beds or layers within thicker sandstone beds are deformed plastically (“plastic deformation” in Fig. 9). In the core of some folds, s–c-structures developed in mud-to siltstones. In other cases, symmetric cuspate–lobate folds are present in sandstones of the steep limbs of folds, whereas the flat limbs are boudinaged (Fig. 10a,b). In the core of the folds, a foliation parallel to the axial plane is present (Fig. 10c). In continuation of anticlinal hinges into an overlying conglomerate, anastomosing shear zones in the conglomerate are present (Fig. 10a, center).

Faulting and stacking of sandstone beds indicate that shear strength of sandstone was high enough to prevent plastic folding comparable to Fig. 10, but not low enough to prevent multiple faulting in sandstones. Compared to folds described up to this point, lithification is more advanced, but not complete, and local plastic folding gives evidence of inhomogeneous lithification. Compared to fluidization, a younger relative age for the folds described here can be deduced due to refolding of fluidized layer 1 in Fig. 6.

2.2.2. Dekametric folds

On a hectometric scale, axial planes of folds in the northern face of Pleiskopf can be traced across 200 m of the sedimentary succession (Fig. 11). Where these large folds are present, the heterolithic units between the thick conglomerate beds show chevron fold style, indicating folding by flexural slip. However, the hinges of folds within conglomerates are thickened, indicating that material flowed toward the hinges. The contacts between
the heterolithic unit and conglomerates at the inner bends of the large folds show metric mullions comparable to those of Fig. 7b, but the sandstones in the heterolithic unit do not show any evidence of plastic deformation.

The dekametric folds described here are part of the tight fold train at the southern margin of the Muttekopf Gosau outcrop. As can be seen in Fig. 3, these folds formed above the tip lines of minor thrusts reaching into the sedimentary succession. During folding, the sandstones were completely lithified, but the conglomerates still behaved plastically.

2.2.3. Normal faults

Synsedimentary faults are a widespread feature in the Gosau Group of the Muttekopf area and occur at all scales of observation, from the hectometric to

Fig. 8. Hydroplastic deformation in a conglomerate bed. A heterolithic unit is repeatedly isoclinally folded into a conglomerate bed. To the right, the lateral margin of the gliding unit is characterized by conical folds, where the heterolithic unit was dragged at the slump edge. Movement of the slump is toward SE, away from the observer. a: poles to bedding of the folded heterolithic unit and the calculated fold axis. b: axial planes of folds within the heterolithic unit. c: fold axes in the heterolithic unit, represented also by rose diagram. The squares record the changing orientation of the sheath fold axis of the fold in the center of the figure. d: poles to bedding planes and calculated fold axis in the zone of conical folds at the lateral margin of the slump.
centimetric scale. Some of the major faults can be seen in Fig. 3 and in Fig. 11. Typically, offset across the major faults decreases upsection and downsection. Offsets of up to 100 m were observed. Thickness changes of sedimentary units occur across these faults, but breccias have never been found to be associated with such faults. Fault orientations are both parallel to the regional fold axis and perpendicular to it.

As the faults lack breccias and only thickness variations are found, they probably never created a major fault scarp. Any surficial depression was immediately filled by sediment. The orientation of the faults is generally parallel to the interpreted dip of the slope, with the hangingwall moving down in the downslope direction.

2.2.4. Downslope gliding of sediment packages

The only possibility to document downslope gliding of sediment packages is to find locally developed ramps at the base of the gliding units (Fig. 12). In most areas, the units move on bedding planes in heterolithic units, and these glide planes are difficult to recognize. In the example shown in Fig. 12, a ramp is developed at the base of a gliding unit, stacking a conglomerate bed. The coarse-grained unit on the top of the translated heterolithic unit fills the topography created by gliding. The geometry of the ramp at the base of the gliding unit suggests NE-directed movement of the hanging wall. In the example shown in Fig. 12, gliding was active under a sediment stack of 6 to 7 m thickness.

3. Discussion

3.1. Slump folds versus tectonic folding

3.1.1. Kinematic model for slumps

Farrell (1984) and Farrell and Eaton (1987) have proposed a dislocation model for the movement of slump sheets. According to the model, movement and deformation of a surficial slump sheet on a basal failure of finite size is spatially organized (Fig. 13a), assuming continuous strain and a layered plastic or brittle material having beds with differing competence which get folded. The lower part of the slump gets shortened, while the upper part of the slump gets extended. Initial deformation in the lower half of the slump will cause bedding-parallel shortening creating fold axes perpendicular to the downslope direction, whereas extension in the upper part of the slump causes boudinage and normal faulting. Further gliding will amplify the existing folds and rotate them progressively toward the downslope direction. When the basal failure propagates downslope, new folds with axes parallel to the tip lines of the glide plane will form. Strain overprinting is achieved by arresting slump movement at the toe and shifting the area of active thickening upslump into the area of former extension, or by arresting slump movement at the head and shifting the area of active extension downslump into the area of former shortening. In both cases, complex strain histories will be restricted to the central part of the slump, where the basal glide plane has maximum offset.

In the dislocation model the orientation of fold axes within the slump body is a function of the maximum offset across the basal glide plane relative to the size of the glide plane. Topography of natural examples of surficial slumps is in accordance with the dislocation model on small and large scale (e.g., Lajoie, 1972; van Weering et al., 1998; Bøe et al., 2000). In the example documented by Lajoie (1972), folds in the shortened part of the slump are of dezimetric scale, whereas the maximum offset across the basal glide plane is about 10 m and therefore large compared to the dimension of the folds. Most of the fold axes are parallel to the dip of slope (Fig. 13b). When measuring
slumpsheet fold axis orientations in outcrop, about 50% of
the folded area will have fold axes parallel to slope dip
(area shaded grey in Fig. 13), 25% parallel to slope strike
(area shaded light grey in Fig. 13), and the remaining 25%
of the area would have an intermediate fold axis orienta-
tion. In slumps with small offsets across the basal glide
plane compared to the dimension of folds, fold axes
should be parallel to slope strike.

The model outlined above describes the geometry of
the distribution of fold axes within slumps, but does not
offer an explanation for the development of isoclinal folds
with axial planes parallel to bedding. Previous workers
have described bedding-parallel folds that are often bou-
dinaged and resemble intrafolial folds (e.g. Woodcock,
1976; Blewett, 1991), and folds with strongly curved
hinges (e.g. Williams et al., 1969; Farrell, 1984; Tobisch,
1984; Farrell and Eaton, 1987). These fold geometries are
frequently observed in shear zones and metamorphic
rocks with mylonitic foliation (e.g. Ramsay and Huber,
1987). Fold axes in such rocks are commonly parallel to
the transport direction of the hanging wall (e.g. Cloos,
1946; Flinn, 1962; Grujic and Mancktelow, 1995). The
reason for this is the tendency of linear elements to rotate
toward parallelism with the transport direction if offset
becomes high in relation to the size of the material line
(e.g. Cobbold and Quinquis, 1980). The occurrence of
folds with geometries indicative for high strain such as
sheath folds in distinct layers suggests a shear zone model
is useful for some of the features observed in slump sheets.

3.1.2. Folding due to tectonic forces

Medium-to small-scale folds in the brittle crust related
to tectonic deformation are either related to a fault or shear
zone, such as fault-propagation folds or fault-bend folds
(e.g. Suppé, 1985), or they are parasitic folds related to
large scale folding. In the first case, folds related to a
single phase of deformation should be coaxial and verge
in one direction. Applying the dislocation model outlined
above, tectonic deformation can be described as defor-
mation related to dislocation of almost infinite size. In the
second case, folds are also coaxial, with fold vergence
having a systematic relationship to large-scale folds. Most
folds related to tectonic deformation affect thick sediment
packages and are not restricted to single layers of rocks.
3.1.3. Relation of lithification to deformation

The main difference between slump-related deformation and tectonic deformation is that the first takes place while the sedimentation process is ongoing, but cannot continue under a thick sedimentary cover, while the second process is persistent through time, at least in areas with ongoing folding. Therefore, soft-sediment deformation due to slumping should be restricted to the initial stages of lithification of a rock, while soft-sediment deformation due to tectonic forces should take place during all stages of lithification.

3.1.4. Interpretation of Upper Gosau folds

The criteria presented above suggest deformation features in the Upper Gosau Subgroup are related to (1) fluid escape and injection driven by pore pressure instability and sediment loading intrinsic to sediment gravity flow deposition, (2) gravity driven glide processes or (3) tectonic deformation. Deformation features closely related to sedimentation are the load structures described in Section 2.2.1 and gliding of sediment packages associated with folding. In both cases, a local topography created by the deformational process was filled by subsequent sedimentation.

Because the size of fluidized units (Section 2.1.2) cannot be infinite, movement of hydroplastically deforming beds which are not disrupted into very small parts within a fluidized matrix should in some way conform with the model for surficial slumps. However, total translation of material is limited, as the fluidized state can not be maintained for a long time (Lowe, 1976). Free movement and chaotic mixing during the fluidization event is not possible if the fluidized layers has abundant hydroplastically deforming “rigid” beds that are in contact with each other. The dominant orientation of bedding is inclined downslope or upslope and fold axes are scattered around an axis perpendicular to the downslope direction (fluidized layer 1 of Fig. 6). This corresponds to the initial stages of downslope movement of a surficial slump in the dislocation model. Occurrence of refolded isoclinal parallel folds might be a consequence of folding in an environment of extreme competence contrasts between the fluidized conglomerate matrix and the cohesive sandstone beds.

Hydroplastic deformation in single or several layers is related to deformation of conglomerate beds (Sections 2.1.3 and 2.1.4). A glide plane parallel to bedding is found in a heterolithic unit directly below a deformed conglomerate bed. Upsection from the glide plane, simple shear affects the thin heterolithic layer and the base of the conglomerate (Figs. 7 and 8). The thickness of the sediment pile transported on top of the glide plane is not known. In these examples, the shear zone model should be applied. Probably localization of gliding just below the coarse-grained
bed is related to loading after deposition of the conglomerate bed. In the case of the asymmetric mullions (Section 2.1.3), finite strains within the shear zone were small, and fold axes parallel to transport. The mullions can probably be seen as an initial stage leading to the pervasive hydroplastic deformation of coarse-grained layers.

All types of deformation discussed in this section up to this point affect single or several layers, and are characterized by hydroplastic folding of sandstone and mudstone beds. No deformations in single or several beds characterized by semi-brittle or brittle deformation were observed. In contrast, soft-sediment folds affecting many beds show plastic deformation (Fig. 10), semi-brittle deformation (Fig. 9) and brittle deformation (Fig. 11, Section 2.2.2) of sandstone. All of the folds described in Section 2.2.2 are coaxial and face N. The verging dekametric folds described in Section 2.2.2 are considered to be of tectonic origin, since they are related to detachments that cut upsection (Fig. 3). This is further supported by verging metric folds with the same geometry that formed during all stages of lithification (see above).

3.2. Relation of structures to sediment transport direction and slope dip

At the northern limb of the main syncline, sediment transport directions are to the WSW, which is parallel to the main axial trend of the large scale folds in the area, or SE, down the northern limb of the main syncline (Fig. 14a). Slump folds indicate SSE-directed gliding of sediment (Figs. 6, 8, 13b), which is down the northern limb of the main syncline. Accordingly, synsedimentary normal faults in the area downthrow the southern block, consistent with a SSE-dipping of the slope (Figs. 3, 9). In the zone of complex folding south of the core of the main syncline, the pattern of sediment transport is complex and locally suggests transverse transport down the limbs of folds, perpendicular to anticlinal hinges. Vergence of slump folds in the area is in some cases opposite to sediment transport (compare flute casts of Fig. 7a and vergence of folds of Fig. 7c,e,f), which suggests some interaction with fold growth. However, the style of deformation does not change along one bed across the hinge of the hectometric Alpjoch syncline (Fig. 7d).

Slope dip of turbidite fans is commonly small, ranging from 1° to 14° (Stow et al., 1996), and in the Muttekopf sediment gravity flow system single conglomerate layers fill topography created by folding (e.g. bed marked with “1” in Figs. 2 and 7d). A change in slope dip to the opposite direction should be connected to a change in the boundary conditions of folding, e.g. the end of growth of

Fig. 13. Models for the movement mechanism of slumps. a: the dislocation model (redrawn from Farrell (1984). b: Fault and fold orientations in a snow slump, redrawn and modified from Lajoie (1972). Bottom right rose diagram for all fold axis orientation in the slump. Grey shading denotes area with fold axes parallel to transport direction, light grey shading area with fold axes parallel to slope strike c: Relation between large scale features of a slump described by the dislocation model and the shear zone at the base of a slump.
3.3. Distribution of deformation structures

The different types of deformational structures described are not evenly distributed. Some of the features like loading structures (Section 2.1.1), fluidization (Section 2.1.2), normal faulting (Section 2.2.3) and down-slope gliding of sediment packages are found all throughout the study area. Other types of deformation are restricted to the area of complex folding in the southeastern part of the Muttekopf outcrop, with only meter-scale tectonic folds present in a few places further west (Fig. 14b). This suggests that slump processes were enhanced where wavelengths and amplitudes of growing folds were smaller. Assuming constant rate of shortening, smaller structures have higher rates of tilting per time unit than large structures. More work is needed to exactly define the relationship between growing folds and slump processes.

4. Conclusions

In the Upper Gosau Subgroup deformation occurred across a broad spectrum of timing and rheology. It is possible to distinguish gravity-driven deformation from tectonic deformation related to folding and thrusting; however, it is not possible to make this distinction from a single outcrop. In the example described in this paper, following arguments are used to separate tectonic and slump structures:

1) Fold axes in slump deposits and fluidized deposits scatter strongly about the transport direction of the unit or the strike of the slope. If folds are not isoclinal, they commonly face downslope. In contrast, axes of folds related to tectonic deformation tightly cluster in a direction perpendicular to tectonic
transport. All folds are asymmetric and consistently face in transport direction.

2) During slumping, shortening is commonly associated with extensional deformation, whereas tectonic folding is not.

3) Whereas slump folds form during the earliest stages of lithification, folds related to tectonic shortening form throughout lithification. Tectonic folds therefore display a much wider range of structures related to folding.

Two models are used to describe slump movement in this paper: First, the dislocation model proposed by Farrell (1984) and Farrell and Eaton (1987), which describes the overall distribution of deformation in a slump. If the slump folds are present at the base of a much larger gliding unit, a shear zone model (e.g. Cobbold and Quinquis, 1980) may be more appropriate. The shear zone model accounts for both the formation of shear folds, and the rotation of axial planes and fold axes of preexisting folds into parallelism with foliation and transport direction, respectively.

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