STRUCTURAL CONTROL OF MULTI-SCALE DISCONTINUITIES ON SLOPE INSTABILITIES
IN CRYSSTALLINE ROCK (MATTER VALLEY, SWITZERLAND)

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Abstract

Rock slope instabilities are considered a major hazard and a limiting factor for the expansion of human settlements in mountainous regions. Structural configuration and slope geometry govern the strength and stability of fractured rock masses in alpine slopes. The western flank of the Matter valley in the Swiss Alps, between the localities of Randa and Sankt Niklaus offers a unique opportunity to improve the understanding of the relationships between rock discontinuities and rock slope failure phenomena. This section of the Matter Valley shows one of the highest topographic gradients in the Alps, and high frequency of landslides. Important data for the present study was collected from field mapping, analysis of aerial- and ground-based remote sensing imagery, digital elevation models (DEM), and displacements from GPS, InSAR and geodetical monitoring.

Foliation and non-penetrative discontinuities were systematically characterized for the entire study area. Discontinuity characterization was mainly based on discontinuity orientation. The smooth continuous distribution of foliation orientation (schistosity in ortho- and para-gneisses) was interpolated from 765 field measurements. Smooth changes occur in the study area from S to N and from bottom to top of the rock slopes. Dip direction values vary from S to NW (180°-305°) and dip angle from gentle to slightly steep values (12° - 35°). Conformity between topography and foliation shows that the slopes are predominantly orthoclinal (49%) and anaclinal (47%) with steepened escarpments. Overdip slopes occupy only 4% of the study area but are important in terms of slope stability. They occur on slopes dipping to the S and SW in the southern part of the area in side valleys below 2000 m asl.

Non-pervasive discontinuities in the area are joints and brittle faults which were arbitrarily classified into meso- and large-scale based on discontinuity length. One meso-scale set and one large-scale set (mostly brittle-ductile shear zones) are parallel to foliation.

For the analysis of meso-scale discontinuity properties, the study area was subdivided into 10 spatial domains. A new GIS based approach was developed to extract large numbers of fracture orientations from a 2.5 m digital elevation model (DEM). The reliability of the results was tested by comparison with field data (2110 fracture measurements) using ANOVA tests, which show that only for 1 fracture set in 1 structural domain, the field and DEM analysis differs significantly. Five meso-scale sets were identified in most domains dipping to the N, E and NE. Spatial variation of fracture orientations was observed from S to N. The study area has been divided in two regions with significant differences in meso-scale fracture orientation, limited by the valley of the Blattbach River. Several types of probability density functions for dip and dip direction were tested to fit to the
fracture set samples in order to identify the best-fit distribution types in all spatial domains. Kolmogorov-Smirnov and Anderson-Darling tests give to Normal, Weibull and Gamma distributions the highest scores.

Four sets of large-scale discontinuities corresponding to brittle faults, fracture zones and brittle-ductile shear zones were identified in the study area based on field measurements (226 values) and DEM analysis of 350 lineaments traces using a Matlab multiple point least squares regression script. DEM-derived fault orientations show a significantly higher scatter than field based data.

An inventory map of rock slope instability phenomena was produced for the area. Terrains were classified by type, age, and magnitude of current rock slope instabilities. The study area can be subdivided in northern and southern regions separated by the Blattbach River. These two domains have different geomorphological evolution expressed by differences in the frequency, depth and width of gullies eroded into the steep sidewalls of the Matter Valley. In the southern domain, many (lateglacial) release surfaces have been found which are mostly located on the front walls of the main valley. In this domain, magnitudes of all type of deposits are comparatively small. In the northern domain, magnitudes of rockfall deposits are larger than in the south, being in the order of several tens of thousands of square meters. Also slightly larger debris flow cones exist in the northern domain. For the northern region most (postglacial) slope failures with genetically correlated deposits are located in the steep walls of the main valley slopes and few in the southwestern facing flanks of the lateral valleys.

In addition, three of the larger current instabilities were investigated in detail. Results show that slope geometry and discontinuity orientation have a primary control on the failure mechanisms. E-NE facing slopes are prone to fail by sliding (northern compartment Walkerschmatt, Sibulbodme), or slumping (Sibulbodme) whereas SE-S facing slopes are more susceptible to fail by toppling (southern compartment Walkerschmatt, Medji rockslide 2002). In most cases, internal deformations of the rock masses are controlled by steep meso-scale fracture sets, and large discontinuities (faults) control the boundaries of the instabilities (e.g. Sibulbodme, Medji, Randa).

Small size rockfall distribution has been assessed by applying physically-based methods for three classic failure mechanisms: planar and wedge sliding, and toppling. Monte Carlo simulations of fracture distributions have been applied in a geographic information system (using ArcGIS) to quantify probability of failure (Pf) based on principles of kinematic analysis. Normal distribution parameters (mean and standard deviation) for orientation and friction angle of meso-scale discontinuities were used. Areas with simulated Pf ≥ 0.2 were taken as potentially active. Simulated areas with high Pf values coincide well with slopes showing high density of former small size rockfall events. For the case of the current large instabilities, the failure mechanisms showing
high Pf values coincide with the observed controlling failure mechanisms. This new simulation approach has a large potential for practical applications.
Zusammenfassung


Effektive Diskontinuitäten beinhalten Klüfte und spröde Verwerfungen, welche entsprechend ihrer Länge als mittel- und grossskalig klassifiziert wurden. Ein mittelskaliges (1 cm-10 m) und ein grossskaliges (> 10 m) Set (hauptsächlich spröd-duktile Scherzonen) liegen parallel zur Foliation. Für die Analyse der Eigenschaften mittelskaliger Diskontinuitäten (hauptsächlich Klüfte) wurde das Studiengebiet in 10 strukturelle Bereiche unterteilt. Um die grosse Anzahl Kluforientierungen aus einem 2,5 m DHM zu extrahieren, wurde ein neuer GIS-basierter Ansatz entwickelt. Die Zuverlässigkeit der Resultate wurde durch einen Vergleich mit Felddaten (2110 Messungen) mithilfe eines ANOVA Tests ermittelt. Nur bei einem Klufset in nur einem strukturellen Bereich wichen Feldmessungen und Resultate der DHM Analyse signifikant voneinander ab. Fünf mittelskalige Sets,

Für das gesamte Studiengebiet wurde eine Inventarkarte für Fels-Instabilitäten hergestellt. Das Gelände wurde anhand des Typs, des Alters und der Magnitude der aktuellen Instabilitäten klassifiziert. Das Gebiet wurde wiederum in ein nördliches und ein südliches Gebiet, getrennt durch den Blattbach, unterteilt. Diese zwei Bereiche haben eine unterschiedliche geomorphologische Evolution erfahren, was sich durch die Häufigkeit, der Tiefe und der Breite der Seitentäler, welche in die steilen Talflanken des Mattertales erodiert wurden, zeigt.

Im südlichen Bereich, hauptsächlich entlang der Flanke des Haupttales, konnten zahlreiche (spätglaziale) Ablösungsflächen beobachtet werden. In diesem Bereich ist die Fläche jeglicher Ablagerungen vergleichsweise klein. Im nördlichen Bereich ist die Fläche der Felssturzablagerungen grösser als im Süden und liegt in der Grössenordnung von 10'000 m2. Ebenfalls leicht grösser sind im nördlichen Bereich die Murgangkegel. Die meisten (postglazialen) Anriss mit genetisch korrelierten Ablagerungen sind im nördlichen Bereich an den steilen Wänden des Haupttales, sowie einige an den südwestlichen Flanken der Seiten, zu finden.

Drei der grösseren Instabilitäten wurden zusätzlich detailliert untersucht. Die Resultate zeigen, dass die Hanggeometrie und Orientierung der Diskontinuitäten die Hauptkontrolle auf den Versagensmechanismus ausüben. Ost bis Nordost gerichtete Hänge neigen zu planarem Gleiten (nördliches Kompartement von Walkerschmatt und Sibulbodme), oder zu rotationellem Gleiten (Slumping; Sibulbodme). Südost bis Süd gerichtete Hänge dagegen haben eine grössere Anfälligkeit zum Versagen als Hakenwurf (südliches Kompartiment von Walkerschmatt und Medji Felssturz von 2002). In den meisten Fällen wird die interne Deformation durch steile mittelskalige Kluftsets
kontrolliert, während grossskalige Diskontinuitäten (Verwerfungen) eher die Grenzen der Instabilitäten darstellen (z.B. Sibulbodme, Medji, Randa).

Chapter 1

Introduction

Large instabilities in natural rock slopes have been always considered as a threatening factor to the development and expansion of human settlements and vital networks and have caused substantial economical and life losses along historical times. In the last decades large slides have occurred in the world (e.g.: Cerro Huascaran in Peru, Vaiont in Italy), which have demonstrated the need for a more comprehensive approach to the mechanisms as well as to the evolution of unstable rock masses. In Switzerland, more than 6% of its territory is prone to landslides (Noverraz & Bonnard, 1990 in Leroi, 2005). Large-scale slides (e.g.: Randa, Goldau, Elm) have shown the capital importance of a deeper understanding of their failure mechanisms to improve the assessment and the forecasting of future events.

Rock slope instabilities on crystalline rocks have predominantly a structural control. Rock discontinuities control the behavior and define the driving mechanisms of the rock slope instabilities at different scales. Structural configuration and geometry govern the deformability, strength and permeability of rock masses. The western flank of the Matter valley in the Swiss Alps, between the localities of Randa and Sankt Niklaus has been chosen for the present study because it offers a great opportunity to improve the understanding of the relationships between rock discontinuities and failure phenomena. The area presents several rock slope instabilities at different scales and allows the acquisition of complete records of discontinuity characteristics due to large rock exposures.

Within the framework of the present project three main activities were carried out: i) structural characterization of the rock slopes based mainly on discontinuity orientation; ii) identification and description of rock slope instabilities at multiple scales, and iii) susceptibility assessment of small size rock fall events using mechanically based models. For all activities data was collected from intensive field work, aerial- and ground-based remote sensing imagery analysis, digital elevation models (DEMs) analysis and geodetical monitoring.

1.1 Structure of the thesis

The present work has been written in a paper format. Each chapter includes its own introduction (with literature review), results, discussion and/or conclusions. Chapter 2, 3, and 4 correspond to the description to the three main activities carried out for this project, which were enunciated above.
Chapter 2 refers to the structural characterization of the natural rock slopes in the study area. Foliation and non-penetrative discontinuities (fractures at meter to kilometer scales) were systematically characterized for the entire study area through the analysis of field data and DEM-based data. The study area is divided in several domains for the analysis.

Chapter 3 investigates the susceptibility of the rock slopes to the occurrence of rock slope instability phenomena in the study area. An inventory map of rock slope instabilities is presented and its different units described. Additionally, three of the larger current rock slope instabilities in the area are analyzed. Finally, the influence of slope geometry and fracture characteristics on the development of the observed landslide phenomena is discussed.

Chapter 4 investigates small size rockfall distribution. Stochastic simulations based on physically based methods are used to assess probability of failure of rock slopes in the area. Three classic rock slope failure mechanism are analyzed: Planar sliding, wedge sliding and toppling based on the principles of kinematic analysis.

Finally, chapter 5 summarizes the work and main outcomes of the thesis.
Chapter 2

Multi-scale rock mass structure from integrated field and high-resolution DEM investigation using Geographic Information Systems (Matter Valley, Switzerland)

2.1 Introduction

2.1.1 Rock Mass Characterization

Geotechnical characterization of rock masses is a common task that has been applied for multiple purposes related with engineering problems such as underground excavations or slope stability assessments. A good knowledge of the internal geometry of a rock mass is essential for the understanding of the behavior when external conditions change or undergo periodical variations (i.e. seasonal changes in environmental conditions). There are several components that can be included to characterize a rock mass and they will depend on the purpose of the characterization. For example, several rock mass characterization systems have been developed to predict rock mass behavior on underground and slope excavation projects. Some of the most popular systems are Q (Barton et al., created on 1974, revised on 1993 and 2002), RMR (Bienawski, created on 1973, revised 1988), and GSI (Hoek, created on 1995, several revisions e.g. Cai et al., 2004). They include mechanical characteristics of the intact rock, physical and geometrical properties of discontinuities and external factors like stress conditions, or groundwater pressure. However, their application requires detailed information of site conditions. This level of detail cannot be achieved on areas of difficult access such as mountainous regions, where characterization of rock masses are carried out involving fewer parameters. The present study is focused on the characterization of the discontinuities (mainly fractures) directed to the assessment of failure potential of crystalline rock slopes. Previous works have shown the importance of fracture properties in the occurrence and development of failures in rock slopes (Brideau et al., 2009; Brideau et al., 2006; Mazzoccola & Hudson, 1996; Brady & Brown, 2004). Fractures have an influence on the failure process because they control the kinematics and stability of larger scale rock masses in slopes.
2.1.2 Discontinuities

The term fracture is used for all kinds of effective discontinuities with zero or low tensile strength. Fractures are classified according with their length and type of movement. Faults are brittle shear zones for which visible displacements have occurred, primarily parallel to the fault plane. Joints are fractures in meso-scale dimension for which no shear offset or dilation is detectable in the field. Schistosity is a penetrative rock fabric and a potential discontinuity, defined by a parallel alignment of platy minerals in metamorphic rocks, producing an inter-leaving or foliation structure.

Fractures have been divided for the present work according to their persistence in large scale discontinuities (visible trace length of hectometers, i.e. more than one hundred meters) and meso-scale discontinuities (visible trace length of meters to decameters; i.e. less than one hundred meters and more than 10 centimeters). Whereas most meso-scale discontinuities are joints, large scale discontinuities are brittle and brittle-ductile shear zones (i.e. faults). The term fracture is used in this paper to describe all types of effective discontinuities (breaks in rocks), irrespectively of size or mode of failure.

Discontinuity orientation has been used in the present analysis as the main discontinuity property and intensive data collection using different methods has been carried out. Orientation is defined by the dip of the line of maximum declination on the discontinuity surface measured from the horizontal, and the dip direction or azimuth of this line, measured clockwise from true north (Brady & Brown, 2004).

Large scale discontinuities play an important role in the development of large instabilities in mountainous environments. Usually, boundaries of deep seated slope gravitational deformations are reactivated pre-existent brittle faults (Agliardi et al., 2001; Brideau et al., 2009). In the study area, the 1991 rockslides at Randa had an obvious structural control as the failed rock mass was segmented in several compartments limited by persistent fractures (Sartori et al., 2003). Moreover, faults control the kinematics of the current instability at Randa (Willenberg et al., 2008; Gischig et al., 2009).

Meso-scale fractures are an important factor that condition smaller scale slope instabilities. For example, the rock slope failure of Medji in 2002 was controlled by two meso-scale joint sets (Ladner et al., 2004). Also the current instability at Calchofen is controlled by meso-scale fractures (Rouiller et al., 1998).
2.1.3 Objectives of Study

Mountainous areas are regions considered prone to the occurrence of catastrophic rock slope failure because of their geomorphological characteristics and discontinuity geometry. The latter is considered an important factor on the occurrence of rock slope instabilities. Large instabilities in natural rock slopes have always represented a threatening factor to the development and expansion of human settlements and vital transportation networks. They have caused substantial economical and life losses during historical time. A better understanding of failure mechanisms is especially important for unstable slopes in fractured rock masses, where failure has been postulated to develop progressively by strength degradation, failure of intact rock bridges, and internal rock mass deformation. Factors controlling the long term evolution of failure surfaces in brittle rock slopes may include cyclic pore pressure and temperature variations, chemical weathering, seismic loading, and slope toe erosion. However, the most important factors that make a slope prone to failure are their structural configuration and geometry, since both together govern the deformability, strength and permeability of any rock mass.

A detailed structural characterization of rock masses is essential for a rock slope stability analysis. For the case of natural rock slopes, because of their accessibility and altitudinal changes, a throughout data collection by traditional field techniques presents several limitations due to spatial coverage and the size of the sampling areas. New techniques for structural characterization of rock masses focused primarily on discontinuity orientation have been developed on recent years. Several of those techniques take advantage of the recent developments on remote sensing data acquisition and digital elevation model generation. New technologies such as light detection and ranging (LIDAR) scanners or digital ground based photogrammetry have greatly contributed to the development of high resolution digital elevation models where new methods for manual and semi-automatic detection of planes representing discontinuities have been created based on vector geometry, (e.g. COLTOP; Jaboyedoff & Couture, 2003). Co-planarity principles have been used to delimit discontinuity trace areas (Lato et al., 2009). Another approach is through the detection of linear objects that represent discontinuities via remote sensing imagery. Orientation extraction of those lineaments is done using vector geometry with surface deformation extracted from digital elevation models in a similar fashion as described on Feng et al. (2001) for the case of linear features surveyed by a non-reflector total station. These works have been focused mainly to specific sites such as open pit mines or selected natural rock slopes. Regional analysis using similar tools are scarce and usually only including one of those considerations.
On the present work, three different approaches for the use of GIS-based tools and DEM analysis on structural characterization of rock masses are presented. They were developed based on the criteria mentioned above. Their results were integrated and compared with the results from field work. Commercial software (ArcGIS) has been used for the analysis. GIS programs have shown to be powerful tools which help to integrate and simultaneously analyze several sources and types of information in a user friendly environment. The area selected for the analysis corresponds to the western flank of the Matter Valley in the southern Swiss Alps. The area presents exceptional conditions for the application and validation of the methodologies presented on this work (abundant rock faces controlled by rock discontinuities, scarce or inexistent vegetation covering rock outcrops). Additionally, the area presents intense rockfall activity expressed as large debris cones deposits to the bottom of rock walls (Figure 2.1). Further descriptions of the study area are presented bellow.

A detailed and systematic structural characterization for the study area does not exist. Detail fracture characterizations have been done for few small areas, which were affected by past failures (Joerg, 2008; Willenberg, 2004) or in small areas with high failure potential (Wagner, 1991). The results from these local investigations are summarized in Chapter 4. For the present work intensive field data collection was carried out in order to derive a comprehensive description of the large and meso-scale discontinuity inventory for the whole study area. New GIS-based remote sensing methodologies for data collection were developed to complement the field data sets.

2.2 Geomorphological and geological settings of study area

2.2.1 Geology and Geomorphology

The study area corresponds to the western flank of the Matter Valley between the localities of Randa and St. Niklaus in the southern Swiss Alps (Figure 2.1). The Matter Valley is a long north-south trending valley of the Matter Vispa River in the Swiss Alps and drains into the Rhône Valley. The study area is composed of 11 slope faces separated by secondary creeks draining to the Matter Vispa, from north to south: Grossgufer, Hohebalme, Guggini, Seemate, Saenggini, Walkerschmatt, Medji, Riedji, Chalchofen, Saelli and Sparru (Figure 2.1).

The area is located in the Penninic Siviez Mischabel nappe which is a tectonic subdivision of the Grand St. Bernard nappe (Escher, 1998; Sartori et al., 2006). The Siviez Mischabel nappe is considered as a recumbent fold more than 40 km width (Genier, 2008) emplaced at 41-36 Ma during the Alpine orogeny (Markley, 1998). It comprises basement rocks that are unconformably overlain by units of
greenschist facies rocks, which share the same Tertiary Alpine thermal and kinematic history (Markley, 1999). Basement units comprise gneiss, schist, and amphibolite rocks deformed and metamorphosed to amphibolite facies during the Hercynian orogeny, intruded by a porphyritic alkaline to subalkaline granite of early Permian age (Thélin, 1987), called the Randa orthogneiss. Cover sequences are formed by a metamorphosed Permo-Carboniferous graphitic mica-schist unit at the base, followed by a thick sequence of Triassic quartzites and green schists (meta-sedimentary rocks), and at the top, a Triassic dolomite and limestone sequence (Markley, 1999).

The Randa orthogneiss is derived from a subalkaline porphyritic granite (Bussy et al., 1996). In the Matter Valley it reaches its maximum thickness of approximately 1000 m (Sartori et al., 2006) and covers most of the study area. It is mainly porphyritic on texture with K-feldspar phenocrysts (Genier, 2008), but can display several other facies: microgranitic border facies, apophyses and leucocratic aplites, schlieren, and mafic and xenolithic enclaves. Decametric zones of dark quartz-porphyries are also found. In the Matter Valley at least three bands of leucocratic and porphyritic metagranitoids interlaced with the cover units have been observed and considered to be related to the Randa orthogneiss (Genier, 2008). Regionally, it presents a strong alpine foliation and two lineations; In the study area has a main direction of foliation dipping towards the S and SW, which varies in azimuth and dip with height (Zueger, 2007).

The valley has been molded primarily by the action of glaciers. During the last glacial maximum (LGM) which occurred 18000 to 20000 yr BP (Kelly, 2004), the Matter Valley was part of the main accumulation area in the Swiss Alps with a decreasing ice thickness towards the Rhône valley. In the area of Zermatt, the ice field reached an elevation of 3010 masl. In the Matter valley, the ice elevation ranged between 2550-2700 m asl, 1300 m above its bottom at Randa (Willenberg, 2004). Ice flowed towards north from a central accumulation region at Zermatt. However, not clear evidences such as trim lines were developed into the valley because the variable bedrock lithologies (Kelly, 2004). Glacial release after melting has been estimated to occur around 10000 yr BP (Winstorfer 1977).

Glacial erosion has conditioned the current morphology of the Matter valley. On the western side, front rockwalls parallel to the valley axis, have high dip angles built by glacial erosion and associated paraglacial processes. The steep slopes overprint the original U-shaped valley and are remolded by postglacial erosional and depositional features (Willenberg, 2004). The eastern side slopes of the valley are moderately inclined, and usually foliation-parallel. They are covered with glacial deposits, debris and rockslide material (Schindler & Eisenlohr 1992).

The nature of the response of glacially steepened rock slopes to deglaciation is strongly conditioned by lithology and structure, and in particular, by joint density, and the orientation and
inclination of discontinuities and planes of weakness (e.g. foliation, bedding) relative to the rock face (Augustinus, 1995 in Ballantyne, 2002). Augustinus (1992) demonstrated that steep massive crystalline rock slopes can remain stable, with no evidence for deep seated slope failure following glacier retreat. Thus the nature, scale and timing of paraglacial rock adjustment is strongly conditioned by rock mass strength and joint network, being observed that some deep seated failures may occur centuries or millennia after deglaciation (Ballantyne, 2002). In the study area, the influence of these conditions has been expressed by various types and magnitudes of past and ongoing rock slope instabilities, which will be summarized in the subsequent sections.

2.2.2 Recent slope instabilities

The rockslide in the western slope close to Randa (Figure 2.2) occurred in 2 different stages in April and May 1991 mobilizing a volume of approximately 30 million of cubic meters. These events destructed the main road and rail line along the valley and dammed the Vispa River that resulted in flooding the town of Randa (Eberhardt et al., 2004). These events have been extensively described in the literature (Wagner 1991, Schindler et al., 1993, Schindler & Eisenlohr 1992; Sartori et al., 2003). Several authors have tried to explain the kinematics and failure process applying diverse techniques based either on 2D numerical simulations (Segalini & Giani 2004; Eberhardt et al., 2004) or 3-D geometrical models representing the sequential release of unstable blocks controlled by the mapped discontinuity geometry (Sartori et al., 2003). The hypothesized causes and mechanisms of these two rockslides include glacial valley erosion and stress accumulation, the destruction of rock bridges between pre-existing fractures, water infiltration and chemical weathering leading to a progressive development of stepped failure surfaces. The progressive failure process was visible in several precursory events within the orthogneiss rock face, such as increasing rock fall activity in the 20 years prior to failure (Schindler et al. 1993; Sartori et al., 2003).

Currently, there is a third creeping mass moving towards the valley, located at an altitude between 2000 and 2400 m.a.s.l. at the crown of the two previous slides, having a volume of 3 – 9 million of cubic meters (Loew et al., 2007). This creeping mass has been mapped and monitored since the 1991 rockslope failures in great detail. Within the framework of a large research project, 3 research boreholes were drilled into the unstable rock mass and instrumented with inclinometers, extensometers, pore pressure transducers, and geophones (Willenberg et al. 2008). The instrumentation has been enlarged through the time, including local geodetic networks, climate monitoring, dynamic instrumentations of active fractures, and investigations of the inaccessible parts of the 1991 failure surfaces with remote sensing techniques such as ground-based differential interferometric radar (Gischig et al., 2009; Moore et al. 2010).
These mapping and monitoring campaigns have led to the development of a unique and comprehensive structural and mechanical model of the current instability (Loew, 2007; Gischig et al., 2009) and an in-depth understanding of the mechanisms driving the current slope movements.

The rockslide of Medji (Figure 2.3), a slope to the west of Sankt Niklaus has not been studied as extensively as Randa. However, a good description of the event is given in Ladner et al. (2004). The rockslide occurred in November 2002 and involved 120000 cubic meters of orthogneiss. The slide has a complex surface of rupture consisting of a series of interconnected en echelon fractures that followed the orientation of the main discontinuity sets in the area. Rain and snow melting, were considered as the triggering factors. Chemical weathering caused strength degradation and favored the process initiation. A protection dam built few weeks before the main event reduced drastically the posterior damage caused by the falling rocks. Before the major event, a monitoring network was installed composed of crackmeters, tiltmeters, inclinometers, piezometers and a geodetic monitoring network. Through the interpretation of ground-based radar interferometry and continuous geodetical monitoring, ongoing movement above the area of the 2002 failure has been detected. The current instability is bounded by pre-existent structures.

2.3 Foliation and topographic conformity

2.3.1 Spatial variation of schistosity orientation

For the study area, penetrative structures are represented by foliation. According to Milnes et al. (2006), foliation is a general term for a planar arrangement of small-scale textural, crystallographic and/or structural features in any type of rock. In the case of the Matter Valley, foliation corresponds to schistosity on ortho- and para-gneisses. Foliation can cause significant anisotropy in the deformability and strength of rock masses (Priest, 1993). In the western flank of the Matter Valley the schistosity persistently dips gently to the south-west however slight changes in orientation exist. An important fracture set has developed parallel to this schistosity, becoming a key feature for the predisposition of rock failures. Intensive field work was carried out in the area to evaluate the spatial variation of its orientation. 765 orientation values were collected in the field, trying to cover equally the entire study area (Figure 2.4). With this data set, the area was divided in five domains by visual inspection according with the variations of the orientation values. As shown in Figure 2.4, the southern part of the area until the Blattbach River has two clear domains topographically distributed. The upper area of the slopes, above the first slope shoulder, has a clear south-western direction of dip, whereas the lower part close to the valley bottom, clearly dips towards the south. From the Blattbach River to Medji, the dip direction of foliation becomes regular along the full
length of the slopes with a general orientation to the south-west. The northernmost part of the area (Sparru) presents a mean dip direction to the west but with a higher scattering than the other areas. This part of the area has been divided in two different domains because of a clear change in the dip angle values. The dip in the lower part ranges from 15 to 25 while dip angle in the upper part reach values up to 36 degrees (the maximum value for the whole area). Data from all domains are plotted in lower hemisphere, equal angle stereographic projections from which the main orientation of each domain was calculated (Table 2.1).

Kriging techniques were used to create continuous surfaces to estimate the variation of the orientation of foliation within the study area. Field data was used as input. Two separate grids were created for dip and dip direction and evaluated together (Figure 2.5). Kriging is frequently used to spatially interpolate point measurements in numerous Earth system science applications such as mining or soil physics. Kriging estimates are determined with an inverse distance-weighted average of the known z values from a surrounding set of sampling points (Meentemeyer & Moody, 2000). The kriging weights are derived from a semi-variogram that describes the spatial distributions of the sampled values. A special type of kriging called disjunctive kriging was chosen for the present work because it is a good estimator, most of the times better than linear kriging methods (e.g. ordinary, universal kriging). It reduces kriging variance and exactness of estimation even over anisotropic samples (Yates & Warrick, 1986). On disjunctive kriging, data are transformed to normal distributions, which produce stationary bi-variate normal distributions for all pairs of data. Conditional probabilities that the true values exceed or are less than a specified critical threshold can then be calculated (Webster & Oliver, 1989). Kriging for dip and dip direction was performed using Geographic Information Systems (GIS). The prediction errors for both parameters are showed in Table 2.2.

2.3.2 Foliation parallel fractures

In the study area, a meso-scale fracture set parallel to the foliation has been identified. In fact, most of the orientations for foliation collected during field work correspond to fractures parallel to foliation. The development of a discontinuity set parallel to pre-existing foliation planes have been observed by many authors (e.g. Milnes et al., 2006; Palmström and Singh, 2001). Usually, foliation and fracture sets are not genetically related but because of the mechanical anisotropy caused by the foliation, fractures follow the planar fabric of the intact rock (Milnes, 2006). In the area this fracture set is persistent but its geometrical characteristics vary greatly. Spacing of this set along the area does not follow a clear spatial variation ranging from few centimeters to few meters. Persistence was observed to follow the trend as in the area of Randa (i.e.: mean value ~ 1 m; Willenberg, 2004).
In the study area brittle-ductile shear zones are preferentially following the same orientation as foliation. They were reported to occur in the area of Randa (Willenberg, 2004). Their thickness ranges from few centimeters to several tens of decimeters. They are characterized by a thick band of mylonite-phylonite caused by shearing, highly weathered, and highly eroded, which imply water circulation into fractures of this set. Within the shear zones a second steeper foliation is sometimes observable. As it is mentioned in Willenberg (2004), shear zones are bounded by brittle fractures that show slickensides. Persistence can be higher than 100 m. Brittle fractures are associated to Riedel structures.

2.3.3 Conformity between schistosity and topography

In the area, front walls of the rock slopes are parallel to the trend of the valley axis, steeply dipping to the E. For that condition, foliation (schistosity) dips into the slope. In the case of the lateral valleys, side walls have slope angles that dip preferentially N, S, NW and SW. With these conditions, foliation can dip out of the slope in some areas. Several slope-foliation configurations can be critical in terms of slope stability and it becomes important to determine areas with critical slope stability conditions. In order to assess the potential that foliation and foliation-related discontinuities have to create instability of the rock slopes in the area, a classification presented by Powell (1875) has been used. Powell (1875) introduced a classification that represents the conformity between topography and penetrative discontinuities which have been extensively used in the literature (Sander, 1970; Cruden, 1989 in Meentemeyer & Moody, 2000). Slopes are classified into three main classes: cataclinal slopes (slope dips in same direction as discontinuity), anacinal slopes (slope dips in opposite direction to discontinuity) and orthoclinal slopes (azimuth of slope is perpendicular to azimuth of discontinuity). Cataclinal slopes are subdivided into overdip (slope dip gentler than discontinuity dip), underdip (slope dip steeper than discontinuity dip) and dip (slope dip similar to discontinuity dip). Anacinal slopes are further divided in normal escarpment (slope dip is perpendicular to discontinuity dip), steepened escarpment (slope dip higher and opposite than discontinuity dip) and subdued escarpment (slope dip lower and opposite than discontinuity dip). Meentemeyer & Moody (2000) proposed a topographic/bedding-plane intersection angle index (TOBIA) to classify natural slopes into the referred classes (Figure 2.6). The system is based on the chord length (L) subtended by the angle between the dip angle of foliation (\(\alpha\)) and the dip angle of the slope (\(A\)) on the unit circle. Mathematically expressed as:

\[
L = \sqrt{(\cos \alpha - \cos A)^2 + (\sin \alpha - \sin A)^2}
\]  \hspace{1cm} (2.1)

Where:
If $0 \leq L \leq 0.7654$ slope is cataclinal
If $0.7654 < L \leq 1.8748$ slope is orthoclinal
If $1.8478 < L \leq 2$ slope is anaclinal

Subclasses are defined from the dip direction angle of foliation ($\theta$) and the aspect of the slope ($S$). Thus:

<table>
<thead>
<tr>
<th></th>
<th>Cataclinal</th>
<th>Anaclinal</th>
</tr>
</thead>
<tbody>
<tr>
<td>$-5^\circ \leq \theta - S \leq 5^\circ$</td>
<td>Dip slope.</td>
<td>Normal escarpment</td>
</tr>
<tr>
<td>$\theta - S &gt; 5^\circ$</td>
<td>Underdip slope.</td>
<td>Subdued escarpment</td>
</tr>
<tr>
<td>$\theta - S &lt; -5^\circ$</td>
<td>Overdip slope.</td>
<td>Steepened escarpment</td>
</tr>
</tbody>
</table>

TOBIA values were calculated for the study area using the grids created for dip and dip direction of the foliation and the slope and aspect grids derived from a high resolution digital elevation model (DEM). Topographic parameters were extracted from a 2.5 m (DEM) created by the assemblage of a 2.5 m DEM created by digital photogrammetry techniques with a set of aerial photographs acquired in 2005 and a 2 m LIDAR airborne-based DEM provided by SWISSTOPO. The DEMs were merged into a joint terrain model as point clouds in ArcGIS and a new grid surface was interpolated using natural neighbors. The LIDAR model was preferred for forested steep areas because it has a higher accuracy in such type of terrains. Additionally LIDAR allows vegetation removal. For areas above 2000 m asl, the photogrammetric model was used. Areas above 2000 m asl contain few or none vegetation reducing its influence on the accuracy of the overall model.

<table>
<thead>
<tr>
<th>Domain</th>
<th>Number of measurements</th>
<th>Mean Orientation</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Dip</td>
<td>Dip Direction</td>
</tr>
<tr>
<td>1</td>
<td>67</td>
<td>25</td>
<td>286</td>
</tr>
<tr>
<td>2</td>
<td>81</td>
<td>17</td>
<td>272</td>
</tr>
<tr>
<td>3</td>
<td>130</td>
<td>16</td>
<td>237</td>
</tr>
<tr>
<td>4</td>
<td>193</td>
<td>14</td>
<td>236</td>
</tr>
<tr>
<td>5</td>
<td>280</td>
<td>15</td>
<td>207</td>
</tr>
</tbody>
</table>

*Table 2.1. Mean orientation values for foliation domains based on field data*
Table 2.2. Values of prediction errors for the interpolated surfaces corresponding to dip and dip direction values using disjunctive kriging.

As it can be seen in Table 2.3 and Figure 2.7, slopes in the area are predominantly orthoclinal (48.76%) and anaclinal (47.03%), being the subclass steepened escarpment the most common (44.31%). Together they occupy more than 95% of the area. Orthoclinal slopes are facing mainly to SE and S (Table 2.4), but have an important percentage of slopes dipping to the N, NE and E (38% of their total). Orthoclinal slopes have dip angles varying mainly between 30° and 60°. In a similar way as slope dip direction, orthoclinal slopes have a high scattering for dip angle varying from gentle slopes (<20°) to very steep slopes (>70°). Orthoclinal slopes are evenly distributed along the area with no observable changes.

Table 2.3. Classification of conformity between foliation and slope orientation according to slope angle.

Steepened escarpments dip preferentially to the E (41.5% of their total). There are no steepened escarpment slopes dipping to SW, W and NW. They have a high slope dip scattering varying mainly from 30° to 70°. There are no slopes dipping less than 10°. Steepened slopes are evenly distributed.
along the area. To the south of the area they are mainly located in the upper parts of the slope (above 2000 m asl) or in the southern flank of the secondary streams.

Overdip slopes are not as common as the other two classes but they represent an important issue in terms of stability because dip foliation is gentler than slope dip (can cause planar sliding along the foliation parallel fractures). They occupy around 3.8% of the area and are consistently dipping to the S-SW (95% of their total). They dip preferentially with angles between 30° and 50°. Overdip slopes are not uniformly distributed in the area. They occur more often to the southern part of the area and their occurrence decreases to the north being rare for slopes to the north of Walkerschmatt (Figure 2.7). Overdip slopes are basically located in the lower part of slopes (mainly below 2000 m asl) and are usually forming part of the northern flanks of secondary streams (which usually cross the area from W to E). The other classes have small representativeness in the area.

2.4 Previous regional fracture investigations

2.4.1 St. Niklaus

Roullier et al. (1998) for the area of Chalchofen, reported 3 main discontinuity sets (Table 2.5) and one main subset with a mean persistence bigger than 50 m and spacing in the order of tens of meters: J1 (024/80), J3 (020/50), J4 (110/50) and J4' (110/70). No further descriptions are presented.

Rovina (unpublished work, 2005) presented a total of four fracture sets for the area of the 2002 Medji rockslide (Table 2.6). Set K1 (025/75) is oriented perpendicular to the valley axis and set K2 (110/80) is oriented parallel to the valley axis. Together these sets are responsible of several pillar-like potential failure masses. The other two sets are less representative but still important: K3 (025/50) and K4 (190/70).

Joerg (2008) described in the area surrounding the rockslide of Medji 2 sets of large scale fracture zones (Table 2.5). The first set (ST1), striking NW – SE, dipping steeply (almost vertical) and a second set (ST2), with a mean orientation of 014/76 and thickness ranging from 15 cm to 30 cm, filled with cataclasite. Faults corresponding to these sets were postulated as the ones that control the 2002 Medji rockslide.

Joerg (2008) described two meso-scale fracture sets (Table 2.6), which were reported as joints: K1 and K2. K1 is described as a set striking to NW-SE (mean orientation 037/77), parallel to a large scale fracture set (ST2). K1 is very persistent with a length in the order of up to tens of meters and spacing intervals from 6 to 20 m. Fractures can be open until 1 m. K2 strikes E – W (mean orientation 096/86), parallel to the valley axis. Persistence can be up to several tens of meters and spacing
varies from 1 to 6 m, decreasing to less than 1 m in areas of poor rock mass quality. Quartz and Chlorite have been observed as common fillings for this fracture set. Open fractures can have apertures up to 70 cm.

2.4.2 Herbriggen

Zueger (2007), for the area located between Randa and Herbriggen, described 5 large scale fracture sets (Table 2.5) with persistence of more than 100 m. Set 1 (192/25) is a set formed by shear zones sub parallel to foliation. Set 2 (076/74) presents often slickensides. It is parallel to the valley axis. Set 3 (012/42) is reported as a ductile shear zone set but it is probably a fault set filled by gouge. Set 4 (021/90 – 201/90) corresponds to a sub vertical fracture zone set that can either dip to the NE or the SW. It is not as frequent as the other three sets. Set 5 (034/69), is described as similar to set 4.

Besides large fracture sets, Zueger (2007) identified 5 sets of meso-scale fractures (Table 2.6). A first set, parallel to foliation, with a mean orientation of 224/14. Other characteristics of the set are persistence 2 m to 10 m, spacing 0.5 m to 1.5 m, planar and closed fractures. A second important set has a mean orientation of 073/70 (parallel to the Matter Valley axis). It was observed that these fractures often present slickensides (162/21). Persistence 1 m to 2 m (up to 10 m), spacing 0.2 m to 3 m, aperture 0.1 cm to 1 cm. Other sets are reported to be less important for the area. Set 3 (028/77), persistence 1 m to 2 m (maximum of 10 m in fracture zones), aperture 0.1 cm to 1 cm. Set 4 (110/80) is not well defined for the whole area, persistence 1 m to 2 m, spacing 1 m to 3 m, aperture less than 0.5 m. Set 5 (348/88), highly scattered, persistence 1 m to 2 m, spacing in the order of several meters, aperture 0.1 cm to 1 cm.

<table>
<thead>
<tr>
<th>Conformity Slope aspect</th>
<th>Orthoclinal Dip Slope</th>
<th>Underdip Slope</th>
<th>Overdip Slope</th>
<th>Normal Escarpment</th>
<th>Subdued Escarpment</th>
<th>Steepened Escarpment</th>
<th>TOTAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>north</td>
<td>6.1142</td>
<td>0.0014</td>
<td>0.0012</td>
<td>0.0021</td>
<td>0.0850</td>
<td>0.0245</td>
<td>3.0647</td>
</tr>
<tr>
<td>north - east</td>
<td>4.9673</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.0834</td>
<td>0.0242</td>
<td>3.0376</td>
</tr>
<tr>
<td>east</td>
<td>7.6618</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>1.0148</td>
<td>0.2396</td>
<td>18.3641</td>
</tr>
<tr>
<td>south - east</td>
<td>16.3330</td>
<td>0.0025</td>
<td>-</td>
<td>0.0226</td>
<td>0.4271</td>
<td>0.0884</td>
<td>5.4806</td>
</tr>
<tr>
<td>south</td>
<td>12.1258</td>
<td>0.0675</td>
<td>0.0282</td>
<td>2.2594</td>
<td>0.0150</td>
<td>0.0018</td>
<td>0.0719</td>
</tr>
<tr>
<td>south - west</td>
<td>0.8429</td>
<td>0.1127</td>
<td>0.0875</td>
<td>1.3455</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>west</td>
<td>0.0350</td>
<td>0.0299</td>
<td>0.0530</td>
<td>0.0801</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>north - west</td>
<td>0.6813</td>
<td>0.0118</td>
<td>0.0802</td>
<td>0.0197</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 2.4. Classification of conformity between foliation and slope orientation according to slope aspect.
2.4.3 Randa

Girod (1999) established a different classification of the fracture sets based on the mineralogical origin of fillings and tectonic history (Tables 2.5 and 2.6). His genetic classes are based on observations in the Matter Vispa by-pass tunnel at Randa. He describes five classes of discontinuities. The first class (D1) corresponds to the shear zones parallel to the foliation with high concentrations of calcite, quartz and chlorite. Usually mylonitic layers are related to ductile shearing. The second class (D2) is related to extensional fractures with persistence in the order of meters and tens of meters and variable orientation (062/70, 135/55, 050/60, 080/40) than can correspond to two sets of conjugated fractures. Class D2 presents fillings of quartz and pyrite, and in some cases slickensides. Class three (D3) corresponds to hybrid fractures (shear plus extension) with orientations of 070-080/70, fillings of chlorite, persistence smaller than other sets (5 m). Small en echelon tension fractures are associated to this genetic class, which have been reactivated by shearing close to the area of the Randa rockslide. Joints are grouped in a single class (D4). According with Girod, joints are defined as smooth planes with absence of fillings associated to fragile elastic deformation. The class is described as variable in orientation, planar and without fillings. A fifth class (D5) is assigned to faults with fillings (gouge) with thickness varying from few centimeters to few decimeters and trace lengths up to several hundred meters. According with the gouge thickness of the fault, they have been classified in thick fillings (031/55), thick to fine fillings (026/90) and fine fillings (010-040/25-60 and 340-350/40-60).

Sartori et al. (2003) reported several sets of “persistent fractures” (Table 2.5) obtained from photo-interpretation of images acquired before, during and after the 1991 rockslides. A differentiation between large and meso-scale fractures is not defined, however, all sets are characterized as persistent which may imply they correspond to large scale fractures. Three main steep orthogonal sets (J2: 060/80, J5: 075/70, J6: 140/80) were found. A single fracture from set 6 (170/80) was reported as the northern limit of the 1991 events. Two less steep fractures (J3: 030/30 and J4: 125/45) were as well reported, being a fracture from J3 set the lower bound of the instable mass that caused the catastrophic events in 1991.

Segalini and Giani (2004) presented four main sets of discontinuities (out of ten) from a structural analysis carried out by Wagner (1991) at Grossgufer (The crown of the May 1991 slope failure at Randa). It is not reported if they correspond to large or meso-scale structures. Two sets (070/66, 160/85) are sub-parallel to the slope and dipping out of the slope with higher angle. Another set (320/35) dips into the slope with a similar angle as the other 2 sets. Finally, a last set (112/46) dipping out of the slope with gentler angle was reported has having a significant importance in the development of the sliding plane for the 1991 rockslide events.
### Table 2.5. Large scale discontinuity sets proposed by other authors for the western flank of the Matter Valley.

<table>
<thead>
<tr>
<th>Author</th>
<th>Set 1</th>
<th>Set 2</th>
<th>Set 3</th>
<th>Set 4</th>
<th>Set 5</th>
<th>Set 6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rouiller</td>
<td>$J_3:020/50$</td>
<td>$J_1:024/80$</td>
<td>$J_4:110/50$</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Girod</td>
<td>$D_5:020-40/90$</td>
<td>$D_5:030-040/50$</td>
<td>$D_3:070/10-80$</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Willenberg</td>
<td>$F_3:335/41$</td>
<td></td>
<td>$F_3:088/65$</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Zueger</td>
<td>$S_4:021/90$</td>
<td>$S_3:012/42$</td>
<td>$S_2:076/74$</td>
<td></td>
<td>$S_5:034/69$</td>
<td></td>
</tr>
<tr>
<td>Joerg</td>
<td>$St_1:195/80$</td>
<td>$St_2:014/76$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Willenberg et al. (2008) for the area of Grossgutef and the slopes above, determined 3 large scale fracture sets (Table 2.5) and 8 meso-scale fracture sets (Table 2.6). Large and small scale discontinuities were defined differently than in the present work. Small scale fractures are those whose lengths are in the order of decimeter-centimeter whereas large fractures are the ones whose lengths are larger than ten meters. Based on lithology and topographic elevation along the Randa slope, a division of the area in three structural compartments was carried out. 5 meso-scale fracture sets are steep fractures appearing in all structural compartments, with the exception of one of them (set 4), which is absent in compartment I (the area of the 1991 rockslide crown). Set 1 dips to the NE and is normal to foliation. Sets 2-5 include tectonic and stress release fractures with highly scattered orientations, what is related to the changes in lithology, folding, faulting and complex stress release conditions in a complex 3D topography. The remaining three sets (6-8) of Willenberg et al. (2008) have moderate dip angles and are not present in all structural compartments. Set 6 appears in compartments II and III (above the current instability). Sets 7 and 8 appear in compartment I (current instability).

For the large scale fractures, Willenberg described all sets as faults and fracture zones. Fractures from set F-1 are brittle–ductile shear zones parallel to foliation. F-2 set corresponds to brittle faults, brittle–ductile shear zones and fracture zones dipping N and NW. The NW-dipping major fractures/faults are densely foliated. The F-3 corresponds to faults and fracture zones that strike N–S, parallel to the valley.
2.5 Geometric Properties of Meso-scale Fractures

2.5.1 Field Investigations

For the meso-scale discontinuities in the study area more than 2110 orientations were measured during field recognition. Lineal and spot mapping were used for data collection following the principles presented by Brady & Brown (2004). For lineal mapping a total of 6 scanlines located in the northern part of the area (N of the Blattbach River) were surveyed with a total of 363 measurements. Besides, data collected by Zueger (2007) was included in the analysis for the southern part of the area (a total of 11 scanlines, 490 additional orientations). 990 orientations were collected from spot mapping. Location of spots and scanlines can be seen in Figure 2.8. No subsurface information was available for this work. Despite the extensive field recognition, wide regions of difficult access were not surveyed. The data was plotted on stereonets (equal area, lower hemisphere) and clustered using Fisher distributions (Figure 2.9). Data related with the set of fractures parallel to foliation was not included in the analysis and it is not presented in the spot mapping stereplots. Fisher distribution is a symmetrical spherical distribution widely used for stereographical analysis in structural geology. The area was divided in two different domains according with changes in properties observed during the field recognition. The limit between domains is the Blattbach River (Figure 2.1). Four persistent sets not parallel to foliation were observed in the whole area. They show differences in their characteristics between the southern and northern domains.

<table>
<thead>
<tr>
<th>Author</th>
<th>Set 1</th>
<th>Set 2</th>
<th>Set 3</th>
<th>Set 4</th>
<th>Set 5</th>
<th>Set 6</th>
<th>Others</th>
</tr>
</thead>
<tbody>
<tr>
<td>Girod</td>
<td>D2:050-060/60-70</td>
<td>D2:135/55</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Joerg</td>
<td>K1:037/77</td>
<td>K2:096/86</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 2.6: Meso-scale discontinuity sets proposed by other authors for the western flank of the Matter Valley

Set 1 (J1, Table 2.10 Figure 2.9) dips to the SE with a large scattering on its dip direction (circa 50 degrees of variation) and a steep to sub-vertical slope angle (larger than 60 degrees). Set 1 has a
trace length ranging in the order of decimeters to meters in the southern domain, increasing substantially to the N where fractures corresponding to this set have lengths in the order of several tens of meters (Figure 2.10). To the N, frequency increases causing the creation of slabs parallel to valley axis (Figure 2.10b). Some of the fractures present fillings composed of quartz and calcite with slickensides. Fractures are usually closed or with small apertures (< 2 cm). In the northern domain, aperture increases towards the slope face, especially in areas of high relief where it can reach values in the order of tens of centimeters (Figure 2.10). It is the most frequent set in the northern domain (Fisher concentration about 5%).

Set 2 (J2, Table 2.10, Figure 2.9) dips to the E-NE with a steep angle (larger than 60 degrees). Its length varies from decimeters to meters. It is the most frequent set in the southern domain (Fisher concentration about 6.5%). To the N, its frequency and length notably decrease (Fisher concentration about 2%). Fractures can present chlorite fillings. J2 has been reported by all authors that have worked in the southern domain (Girod, 1999; Willenberg, 2004; Zueger, 2007). However, no references exist for this set in the northern domain showing its low rate of occurrence. Its high frequency in the south of the study area can be explained because it corresponds to 2 parallel sets with different origin. A first set of tectonic origin which present slickensides and chlorite infillings and a second set caused by glacier-retreat related processes (exfoliation joints) which is more densely distributed to the S of the Blattbach River. To the N of the Blattbach River the mean orientation of the front face of the slopes rotates about 15° clockwise. This change in the slope orientation produces a decreasing in the frequency of J2 as it can be observed from the stereographic analysis. To the north of the Blattbach River J1 increases its frequency because exfoliation joints occurred at this orientation (parallel to the valley axis). However, as in the case of J2, a second parallel set of fractures present evidences of tectonic origin (slickensides).

Set 3 (Table 2.10, Figure 2.9) dips towards the NE with steep angles (> 60 degrees). It has a similar frequency in the northern and southern domain (Fisher concentration about 4%). This set strikes perpendicular to the valley axis. It has been observed in both domains with more or less no variations in orientation.

Set 4 (Table 2.10, Figure 2.9) dips towards the NNE-SSW with steep to vertical angles (> 60 degrees), perpendicular to the valley axis. There is a slight variation in its strike between northern and southern domain (ca. 10 degrees). The frequency is higher in the northern domain (Fisher concentration about 4%) than in the southern domain (Fisher concentration about 2%).
2.5.2 DEM and GIS-based fracture investigations

In order to improve the spatial coverage of the locations sampled for fracture data, a DTM and GIS-based method was implemented. The small scale slope morphology in the Matter Valley is strongly controlled by the rock mass structure and rock walls follow discontinuity orientations (Figures 2.10 & 2.13). This fact along with the scarce or missing vegetation cover allows extracting indirectly discontinuity orientations from high resolution digital elevation models (HR-DEM).

Two HR-DEMs were available for this work, created by different methods. A 2.5 m model by digital photogrammetry techniques, created by the group of Photogrammetry and Remote Sensing ETH Zurich, from a set of 62 aerial photographs acquired in August 17th 2005 with orientation data from GPS measured control points, a derived point cloud of around 65 million points, and an accuracy of 0.5 to 1 m. The second model is a 2 m model derived from airborne-based LIDAR data created by SWISSTOPO available for areas up to 2000 m asl with a point spacing of less than 2 m and a vertical point accuracy better than 0.3 m. Because of the way each DEM is created they present different characteristics. Kraus (2007) presents a brief summary of their characteristics. Photogrammetric models perform a geometrical reconstruction of an object in 3D from photographic images, where a photograph defines a bundle of rays or directions. A point on an object is reconstructed when the point is intersected from at least two directions. The directions or rays are generated by natural light and recorded by passive sensors (cameras). For LIDAR models 3D reconstruction is done by measuring objects (or point in objects) from a single location, collecting them as fields of directions and distances (polar coordinates plus azimuths). Position of the device at the moment of collection should be known (GPS/IMU information). Laser scanners are active sensors.

LIDAR based models present a higher accuracy for steep slopes because position is directly measured during data acquisition which brings real information for interpolation. However, accuracy of the model is strongly dependent on point density. For photogrammetry-based models to obtain the same accuracy, a substantial amount of processing is needed by manual addition/correction of control points because the DEM construction is based on image matching and automatic object detection to generate an object-based point cloud. Usually just few points are placed automatically in steep slopes and they are often misplaced. Another significant advantage of using LIDAR models is that they obtain superior results in forested areas as they only need a single location measurement whereas the photogrammetry based models need to match a ground point in at least two images (forested areas are highly textured zones where changes in the position of the observer produce a strong change in the contrast of objects). Moreover, airborne based LIDAR
models can remove forest expression from the models using only backscatter readings of last echoes which allows to build a real ground representation.

From field work it was observed that steep fracture sets often have a representation on the slope walls and that gentle dipping fracture sets represented on rock walls are usually covered by low vegetation (Fig. 2.10). Because of these two reasons, the airborne based LIDAR model was chosen due to its better accuracy on steep slopes and the possibility of having ground representation on forested areas. The influence of those factors in the area is described in a profile comparison of the area of Saenggini shown in Figure 2.11. For the upper part of the slope a difference in height in a relatively flat area is caused by the presence of vegetation (forest). In the photogrammetric model loss of features and flattening of walls are observed in the intermediate part of the slope where steeper walls are located. For the lower part of the slope, a difference in height is observed due to the presence of forest. Differences in height are in the order of 20 to 30 m.

To visualize the changes in slope orientation, a 3D color shaded relief map (Jaboyedoff, 2004) was derived from the LIDAR DEM using ArcGIS (Figure 2.12a). A series of built-in tools were used for its construction. A 3D color shaded relief map is a color-coded image based on HSV (Hue-Saturation-Value) color models showing changes in color according with the changes in slope orientation. In a HSV color model the color space is represented by a single cone (Figure 2.12b). Saturation is represented by the dip angle (slope). Hue corresponds to the dip direction angle (aspect). Value is assumed to be constant and equal to 90. Because ArcGIS cannot build HSV images a transformation to RGB color model was carried out. Figure 2.12c shows a flowchart of the process. The RGB color model is represented by a Cartesian coordinate system with each axis representing a primary color (red, green and blue). In order to get a better visualization of the changes, the color palette was inverted.

A careful selection of the DEM planes used for the fracture analysis was carried out taking in consideration that not all values in the 3D shaded relief image represent fracture orientations. Sampled areas were chosen trying to avoid the presence of artifacts. Presence of thick soil covers and overhanging blocks are the main source of errors. Thick soil covers are present in areas densely forested where last echoes LIDAR data recreate the ground topography but disguised rock exposure morphology (Figure 2.13a). Overhanging blocks hide rock walls to the sensor causing a deficient data collection and a wrong surface representation (Figure 2.13b). Selection of cells was done using 3D visualizations (an orthophoto mosaic created with aerial photographs acquired in 2005 was used as the top-most layer) and photographs of the slopes taken from different angles during the field data collection campaigns. Selected sampling areas (Figures 2.14) were digitized in ArcGIS. Values for dip and dip direction were individually extracted and exported in a matrix-shape ASCII format where
cells outside of the selected areas where assigned a value of zero. The resulting files were converted to single column files in Microsoft Excel (as independent files for dip and dip direction) and exported to Rocscience DIPS to be plotted in stereonets (equal area, lower hemisphere). A total of more than 120,000 orientations were obtained. A source of bias was observed to result from the fact that steeply inclined surfaces are covered by a lower number of pixels than gentle inclined surfaces; a schematic representation is shown in Figure 2.15a. To correct the bias, areal correction was performed in a similar way as Terzaghi’s correction.

Given a plane A, representing a surface with an area $A$ equal to $I \times I$ ($I$ is the pixel size) and inclination angle zero, a plane B with a similar area but an inclination angle equal to $\alpha$, and a plane $B'$ result of the projection of plane B on a flat surface (Figure 2.15b), we obtain:

$$B' = A \times \cos \alpha$$ (2.2)

Assuming $A$ is equal to 1, then

$$W = \frac{1}{\cos \alpha}$$ (2.3)

Where $W$ is the weight assigned to each orientation for correction, assuming that each orientation corresponds to a unit surface (pixel). Each orientation was repeated $W$ number of times on the stereplots in order to correct the bias.

To investigate the detailed spatial variation of the fracture sets, the study area was divided in 10 sub-domains, each sub-domain corresponding to the area occupied by a single slope (Figure 2.1). Fisher distributions were used to observe clusters and group the data into sets. The new data collected helped to improve the spatial distribution of the information as it can be observed in Figure 2.14. A comparison between field data and the DEM derived data was carried out in order to validate the results.

### 2.5.3 Comparison between field and DEM-GIS-based fracture data

For comparison, field data was divided in the same 10 sub-domains as the DEM-GIS-based data and plotted in stereonets to cluster the data using Fisher distributions (Figures 2.16a & 2.16b). Measurements from each cluster were separated in their components (dip and dip direction) for further analysis. To assess the correspondence between data obtained from field work and DEMs, statistical analysis were performed. First, several distribution types for dip and dip direction, recommended in the literature, were compared and fitted to the field and DEM-based data. Second, distribution parameters for the previously chosen distribution were obtained for dip and dip direction of each set in every sub-domain for both field and DEM-based data. Finally, field data and
DEM-based data were compared using statistical hypothesis testing (ANOVA index) in order to evaluate correspondence between the different sources.

2.5.3.1 Analysis of Orientation Distributions

Sixteen different types of distribution functions for dip and dip direction were analyzed in order to identify the best models for the orientation variations of the meso-scale discontinuity sets in the study area. For each sub-domain, fracture orientation data from DEM and field analysis were separated into the 4 inclined sets defined above using the code DIPS (Rocscience). The individual fracture orientations were then separated into their components (dip and dip direction) and evaluated using the program EasyFit, a standard application to fit probability distributions to sample data. The goodness of fit for distributions was evaluated using two different tests: Kolgomorov-Smirnov and Anderson-Darling. Kolmogorov-Smirnov (K-S) tests are used to decide if a sample comes from a population with a specific distribution. It is based on an empirical distribution function and calculates the largest vertical difference between the theoretical and the empirical cumulative distribution function. Anderson-Darling (A-D) tests are a modification of the K-S tests. This type of test gives more weight to the tails than K-S tests and does not depend on the specific distribution being tested. The fits of all sampled distributions were then ranked according to the K-S test and A-D test.

An example of the results for dip direction of J3 is shown in Table 2.7 and Figure 2.17 for the case of Walkerschmatt. This example is typical of the behavior presented for the other samples. Weibull, Normal and Gamma distributions are the distribution types that present the best results for both goodness of fit test K-S and A-D. The Uniform distribution scores well for K-S test but it has high P-values for A-D test because of the histogram following a Gaussian bell shape. Figure 2.18 shows a summary of the scores reached by the distributions for all samples extracted by DEM and field analysis. In total, 78 samples corresponding to ten sub-domains, 4 fracture sets per sub-domain, 2 samples each fracture set (DEM-based and field data samples), except for the case of Riedji where set 2 was not observed. The three best fit distribution types for all 78 samples where used to produce Figure 2.18. As mentioned before for K-S test, Weibull, Uniform, Normal and Gamma distributions have the best fit representing 96% of the total. Weibull (2P and 3P) has a total of 41.5% of all highest scores, followed by Uniform (39.9%, continuous and discrete), Normal (14.1%) and Gamma (10.3%, 2P and 3P). Other distributions are not representative. For A-D tests Gamma distributions are the distribution that best fit the data with a frequency of 37.5%, followed by Normal (25.2%), Weibull (22.2%) and Exponential (7.2%) being in total about 92% of the total.
Normal, Weibull and Gamma distributions are the best ranked for both tests. Normal distribution is consistently ranked in a high position while Weibull scores better for K-S test and Gamma for A-D test. Normal distribution is one of the three best fit distributions for 42.34% of the samples according to K-S test. This value rises to 75.68% for A-D test. Because of these results it was assumed for further analysis that samples follow normal distributions and subsequently their parameters were calculated for both field work and DEM based samples (Tables 2.8 to 11). It has to be mentioned that not all samples include all four reported fracture sets. Set 2 is persistently missed from the field work data samples in the sub-domains to the north of the Blattbach River but appears on the DEM based samples except for the case of Riedji. This fact can be related to the reduction in the frequency of fracture set 2 observed during field work recognition and to the small size of the samples collected during field work in the sub-domains to the north of the Blattbach River due to the difficult access to the area.

<table>
<thead>
<tr>
<th>Distribution</th>
<th>Kolmogorov Smirnov Statistic</th>
<th>Anderson Darling Statistic</th>
<th>Rank</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weibull (3P)</td>
<td>0.04777</td>
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</tr>
<tr>
<td>Normal</td>
<td>0.05015</td>
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<td>1.9497</td>
</tr>
<tr>
<td>Gamma (3P)</td>
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<td>1.8811</td>
</tr>
<tr>
<td>Uniform</td>
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<td>4</td>
<td>16.232</td>
</tr>
<tr>
<td>Weibull</td>
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<td>21.767</td>
</tr>
<tr>
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<td>341.54</td>
</tr>
<tr>
<td>Gamma</td>
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<td>7</td>
<td>3.9937</td>
</tr>
<tr>
<td>Neg. Binomial</td>
<td>0.12985</td>
<td>8</td>
<td>49.539</td>
</tr>
<tr>
<td>Exponential (2P)</td>
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<td>3.2611</td>
</tr>
<tr>
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<td>816.64</td>
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<td></td>
</tr>
<tr>
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</tr>
<tr>
<td>Bernoulli</td>
<td>No fit (data max &gt; 1)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 2.7. Summary of statistics for several distribution types using dip direction set 3 sample acquired by DEM analysis for the area of Walkerschmatt.

2.5.3.2 Equivalence of field and DEM orientation distributions

To evaluate the equivalence between the fracture orientations collected in the field with data extracted from DEM analysis a technique known as the “analysis of variance” (ANOVA) was used. This technique allows to study the homogeneity of a group of samples determining whether sample mean and standard deviations can be considered as equal (null hypothesis) or not (alternate
hypothesis) based on an analysis of the variance within and between samples. The variance is represented by:

\[ SSTot = \sum_{i=1}^{N} (x_i - \bar{X})^2 \]  

(2.4)

Where SSTot corresponds to the sum of between samples squared deviations and within samples squared deviations for a set of \( k \) samples with:

- Sample sizes \( n_1, \ldots, n_k \)
- Sample means \( \bar{X}_1, \ldots, \bar{X}_k \)
- Sample variances \( s_1^2, \ldots, s_k^2 \)
- Total count \( N = \sum_{i=1}^{k} n_i \)

Grand total observations \( T = \sum_{i=1}^{k} n_i \bar{X}_i \)

Grand mean of all observations \( \bar{X} = \frac{T}{N} \)

Alternatively, SSTot can be expressed as:

\[ SSTot = \sum_{i=1}^{k} \eta_i (\bar{X}_j - \bar{X})^2 + \sum_{i=1}^{k} (n_i - 1)s_i^2 = BSS + ISS \]  

(2.5)

Where BSS is the sum of between-sample squared deviations and ISS is the sum of within-sample squared deviations (Borradaile, 2003).

The ANOVA test was carried out for each pair of samples for all fracture sets in all sub-domains (\( k \) value of 2 for two samples for every population). The null hypothesis is set as if samples come from a population with the same mean (\( Ho: \bar{X}_1 = \bar{X}_2 \)). The alternative hypothesis is that their means differ from each other (\( Ha: \bar{X}_1 \neq \bar{X}_2 \)). The test for this case is a ratio between the variances between-samples and within-sample with degrees of freedom \( V_1=1 \) and \( V_2 = N-2 \). The F-distribution for this case is:

\[ F = \frac{BSS}{ISS} \cdot \frac{N-2}{N-2} \]  

(2.6)

F-distribution values were calculated for all cases where data from both sources were available (Table 2.12) and compared with the critical value of 3.84 for a confidence level (\( \alpha \)) of 0.05.
(Borradaile, 2003). 3.84 corresponds to the critical value assigned to samples with $V_{2>120}$. F-distribution values vary between 0.0006 and 5.62. Only in one case Ho is not accomplished (Saenggini dip direction set 4, F-distribution value of 5.62), i.e. only at set 4 at Saenggini the orientation distributions derived in the field and from DEM analysis differ significantly.

<table>
<thead>
<tr>
<th>Sub-domain</th>
<th>Set 1 Dip</th>
<th>Set 1 Direction</th>
<th>Set 2 Dip</th>
<th>Set 2 Direction</th>
<th>Set 3 Dip</th>
<th>Set 3 Direction</th>
<th>Set 4 Dip</th>
<th>Set 4 Direction</th>
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<td>64.08</td>
<td>19.72</td>
<td>81.22</td>
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</tr>
<tr>
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<td>123.17</td>
<td>71.38</td>
<td>75.82</td>
<td>76.11</td>
<td>24.58</td>
<td>80.10</td>
<td>179.02</td>
</tr>
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<td>49.9</td>
<td>14.57</td>
<td>82.03</td>
<td>178.53</td>
</tr>
<tr>
<td>Saenggini</td>
<td>83.82</td>
<td>108.88</td>
<td>77.71</td>
<td>70.80</td>
<td>58.47</td>
<td>37.43</td>
<td>83.36</td>
<td>178.02</td>
</tr>
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<td>65.71</td>
<td>24.85</td>
<td>81.97</td>
<td>192.29</td>
</tr>
<tr>
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<td>65.09</td>
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<td>75.89</td>
<td>47.43</td>
<td>75.58</td>
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</tr>
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<td>-</td>
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<td>35.39</td>
<td>78.89</td>
<td>174.56</td>
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<td>Saelli</td>
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<td>75.54</td>
<td>72.21</td>
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<td>25.88</td>
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</tr>
<tr>
<td>Chalchofen</td>
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<td>70.84</td>
<td>66.47</td>
<td>45.05</td>
<td>20.31</td>
<td>82.93</td>
<td>169.59</td>
</tr>
<tr>
<td>Sparru</td>
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<td>58.63</td>
<td>68.77</td>
<td>62.29</td>
<td>25.95</td>
<td>78.11</td>
<td>170.99</td>
</tr>
</tbody>
</table>

Table 2.8. Mean values for orientation of meso-scale fracture samples extracted by DEM analysis.

Set 1 (J1) dips with values ranging from 63 to 81. A clear decrease in the steepness is observed from S to N. A sharp change is observed at the Blattbach River where dip angle decreases 7 degrees. Set 1 dips towards the SE to the S of the study area and rotates counterclockwise, dipping to the E in the Saenggini sub-domain. From Saenggini to the N, dip direction is about 115º until the study area limits. At Sparru, dip direction rotates towards the SE, getting a similar orientation as in the southern part of the area.

Set 2 (J2) dips with angles between 58 and 71 degrees. There is no well defined trend for the whole area. The lower dip angle values are found to the N of the Blattbach River in Medji and Sparru sub-domains. This fracture set dips toward the NNE for the whole area. A small counterclockwise rotation is observable from S to N. A peak in the dip direction angle value is found at Walkerschmatt, north of the Blattbach River.

Set 3 (J3) dips consistently to the NNE-E with moderate to steep angles. There is not a transitional change in the orientation into the study area. However, there are strong variations from one sub-
domain to the other. Steeper dip angles are found in the central part of the study area (between Saenggini and Riedji).

<table>
<thead>
<tr>
<th>Sub-domain</th>
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<th></th>
<th>Set 2</th>
<th></th>
<th>Set 3</th>
<th></th>
<th>Set 4</th>
<th></th>
</tr>
</thead>
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<tr>
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<td>Direction</td>
<td>Direction</td>
<td>Direction</td>
<td>Direction</td>
<td>Direction</td>
</tr>
<tr>
<td><strong>Hohebalme</strong></td>
<td>4.42</td>
<td>8.85</td>
<td>7.67</td>
<td>10.22</td>
<td>7.96</td>
<td>10.24</td>
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<td>9.91</td>
</tr>
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<td>10.99</td>
<td>6.88</td>
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<td>5.31</td>
<td>5.97</td>
<td>5.55</td>
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<td>8.66</td>
<td>4.12</td>
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<td>11.65</td>
<td>3.08</td>
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<td>12.40</td>
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<td>10.85</td>
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<td>7.98</td>
<td>6.90</td>
<td>11.17</td>
<td>5.88</td>
<td>9.80</td>
</tr>
<tr>
<td><strong>Medji</strong></td>
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<td>5.14</td>
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<td>5.94</td>
<td>7.75</td>
<td>6.91</td>
<td>7.80</td>
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<td><strong>Riedji</strong></td>
<td>3.43</td>
<td>12.97</td>
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<td>6.81</td>
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<td>10.80</td>
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<td>5.98</td>
</tr>
</tbody>
</table>

Table 2.9. Standard deviation values for orientation of meso-scale fracture samples extracted by DEM analysis.

Set 4 (J4) dips steeply to the S having a small clockwise rotation from S to N of the study area (seven degrees). The highest variation on dip direction is observed at the Walkerschmatt sub-domain with a difference of about 15 degrees respect to the main trend. Dip angle values remain similar for the whole study area.

### 2.6 Large Scale Fractures (Faults)

#### 2.6.1 Field Investigations

Four main sets of large scale fractures were found and described during field recognition. For this work 81 orientations of discontinuities collected by Zuger (2007) have been included for the stereoscopic analysis. 145 additional large scale fractures were identified and described during field recognition for the present work. A stereoplot showing fracture orientations is shown in Figure 2.19. The sets present some similarities to descriptions found in the local literature (Section 4). The sets comprise fractures characterized as faults, fracture zones and brittle-ductile shear zones.
Descriptions follow the nomenclature presented by Willenberg et al. (2008). A fault is characterized as a structure with a significant differential displacement. Fracture zones are described as wide areas with intense brittle fracturing but minor differential displacement. Ductile shear zones correspond to densely foliated and fine-grained schists and gneisses (mainly phyllonites and mylonites). Ductile shear zones in the study area often have undergone subsequent brittle deformation and are called brittle-ductile shear zones.

<table>
<thead>
<tr>
<th>Sub-domain</th>
<th>Set J1</th>
<th>Set J2</th>
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<th>Set J4</th>
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<td>Dip</td>
<td>Dip</td>
</tr>
<tr>
<td></td>
<td>Direction</td>
<td>Direction</td>
<td>Direction</td>
<td>Direction</td>
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<td>126.65</td>
<td>77.52</td>
<td>22.56</td>
<td>161.63</td>
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<td>69.41</td>
<td>76.64</td>
<td>82.17</td>
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<td>72.22</td>
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<td>197.0</td>
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<tr>
<td>Riedji</td>
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<td>78.25</td>
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<tr>
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<tr>
<td></td>
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<td>-</td>
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<td>181.0</td>
</tr>
</tbody>
</table>

Table 2.10. Mean values for orientations of meso-scale fractures collected during field work (Subdomains are listed from south to north).

Set 1 (F1; 220/13) corresponds to large scale fractures parallel to foliation (Figure 2.19). This set presents a high level of spatial scattering gradually varying its orientation from S to N. F1 has been described briefly in chapter 4. Set 2 (F2; 010/46) is mainly composed by faults and fracture zones. A specimen of this set has been identified by Sartori et al. (2003) to be the basal plane for the failure of April 1991 at Randa. F2 faults can have traces of several hundred of meters with thick clay gouges and cataclasites (Figure 2.20). Set 3 (F3; 077/68) correspond to faults with slickensides and chlorite infillings that on field have been found to have traces between 5 to 100 m. This set corresponds to set F3o from Gischig et al. (submitted), one of the fractures from this set works as the lateral release plane of the current instability at Randa. Willenberg et al. (2008) argue that fractures of this set can be clearly observed on aerial images with traces that can reach 100m. Most of the larger structures are fracture zones i.e. areas with a high frequency of fractures. This set is parallel to the meso-scale fracture set J2 presented above for the southern part of the study area. Towards the N, fracture
zones with this orientation are less frequent. However, some fracture zones are still observed in the northern region of the area but with a slightly different orientation, following set J1. Persistence and frequency of this set increase towards the N. Set 4 (F4) are steep faults with fine grained gouge infillings (clay) in the paragneiss, and fracture zones following discontinuities of meso-scale set J4 in the orthogneiss. Discontinuities from this set can have lengths in the order of tens to hundreds of meters. Figure 2.21 shows examples of all four sets for the area.

<table>
<thead>
<tr>
<th>Sub-domain</th>
<th>Dip</th>
<th>Dip Direction</th>
<th>Dip</th>
<th>Dip Direction</th>
<th>Dip</th>
<th>Dip Direction</th>
<th>Dip</th>
<th>Dip Direction</th>
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</thead>
<tbody>
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<td>5.06</td>
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<td>6.30</td>
<td>5.53</td>
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<td>73.88</td>
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</table>

Table 2.11. Standard deviation values for orientation of meso-scale fracture samples collected during field work.

### 2.6.2 DEM-based extraction of large scale lineaments

Approximately 650 lineaments that were assumed to correspond to large fractures were identified and mapped during field work from the other side of the valley and digitized in ArcGIS as polylines (Figure 2.21). Other lineaments were added based on interpretation of digital aerial photographs taken in 2005. Lineament traces were modified in order to fit to morphological features observed on the 3D shaded relief map. Lineaments that did not have a morphological expression in the DEM were discarded. All lineaments shorter than 100 m were discarded assuming that they do not correspond to what was defined as large scale discontinuity in the present work. A total of 350 lineaments were kept for further analysis. 3D coordinates for the vertices of the lineaments (x,y,z coordinates) were calculated based on the high resolution digital elevation models described previously and processed with a script written in Matlab to calculate the best fit plane for
each lineament by multiple regression using least squares methods. The code calculates the
direction cosines for the normal vector to the best fit plane based on at least 3 vertices. Using the
direction cosines as input coordinates, dip and dip direction for each lineament were calculated by
vector geometry.

<table>
<thead>
<tr>
<th>Sub-domain</th>
<th>Set 1</th>
<th>Set 2</th>
<th>Set 3</th>
<th>Set 4</th>
</tr>
</thead>
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<td>Dip</td>
</tr>
<tr>
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<td>Direction</td>
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Table 2.12. F-distribution values calculated using the ANOVA technique for the meso-scale fracture sets.

The results show a small clockwise rotation of the F2 fracture set (Figure 2.22) and a reduction of
the dip angles of the steep sets (F3 and F4) with respect to the field data. Rotation varies between 5°
and 15°. Reduction in the dip angle is about 10° for the case of set F3. Only few planes from Set F1 are
observable in the DEM. This can be explained by the fact that they usually do not have a
morphological expression (planes dip into the slope). There is significantly higher scattering than in
the field derived orientation data. This has already been previously reported by other authors (e.g.:
Lato et al., 2009, Sturzenegger & Stead, 2009) and related to the waviness of the fault plane and to
the position of the fracture traces with respect to the position of the laser scanner during
acquisition. Fracture sets parallel to the line of sight of the device present higher errors in their
orientation values. Length of the lineaments plays a role in the quality of the orientation results too.
In spite of all these sources of errors, the three inclined large scale fracture sets could be clearly
recognized and clustered (Figure 2.22). Similar to what has been observed with the meso-scale
fracture sets (Chapter 5), there is a difference between the geometrical characteristics of large scale
fracture sets between the northern and southern areas (Figures 2.23a and 2.23b). A clear limit again
is the Blattbach River. DEM-derived large scale fractures to the south of the Blattbach River present a
spatial distribution comparable to what has been observed during field work. The four main sets can be clearly differentiated. To the north, differences between sets are less obvious. Lineaments representing structures parallel to foliation were not found. Besides the clusters related with the large scale discontinuity sets, there is a concentration of values similar in orientation to the meso-scale fracture set J1 in the northern region. This could be explained by the increasing persistence of J1 fractures to the N of the Blattbach River, leading to trace lengths bigger than hundred meters, which is the cut-off for large scale lineaments in this work. As it can be observed on Figures 2.23a and 2.23b, current morphology of the slopes has a strong structural control. To the area to the N of the Blattbach River large scale structures have a strong influence in the morphological configuration.

<table>
<thead>
<tr>
<th>Sub-domain</th>
<th>Set 1</th>
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<th>Set 4</th>
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</table>

Table 2.13. Summary of mean values for meso-scale fracture set orientations. Black figures correspond to grand mean values and blue figures to only DEM based values.

### 2.7 Discussion

#### 2.7.1 Methodology

The study area of this project shows a very rugged and steep topography which makes field investigations of fracture pattern a time-consuming and risky endeavor. For such areas new applications of GIS-based techniques are presented here. These techniques benefit of the recent development of new methodologies for DEM generation via new remote sensing devices (e.g.
airborne-based LIDAR) and improvements on more traditional techniques (e.g. digital aerial photogrammetry) which allows the generation of more accurate and detailed DEMs (increase on position and resolution). For the application of the techniques presented here, high resolution DEMs and optical images are needed in order to obtain reliable results. This is more evident for the characterization of meso-scale fractures. In areas with high relief such as the Matter Valley, steep slopes are still affected by artifacts e.g. exaggeration on rockwall dip angle caused by overhanging blocks. A manual selection of sample sites is required. It has been done for the present work through field recognition and analysis of optical images (photographs). Comparison between field values and DEM extracted values show high level of correspondence. Results obtained by the application of ANOVA gives a good correlation.

2.7.2 Discontinuity Inventory

Discontinuities are the most important factor for the development of instabilities in hard rocks, especially for the study area in the western flank of the Matter Valley. Both, meso-scale and large scale discontinuities have controlled former past events (e.g.: Randa and Medji).

In the study area foliation is generally expressed as schistosity. Schistosity dips into the slope with clear variations. Schistosity varies along slope height and on a planimetric perspective (from S to N). In the southern half of the area, the spatial variation on its orientation exists from S to N and from bottom to top of the slopes. Close to the valley bottom, foliation dips persistently to the S with angles varying between 190° and 210°. It transitionally rotates clockwise dipping to the SW at the top end of the slopes. The maximum rotation is reached to the southern limit for the Randa (Grossgufer) and Hohebalme slopes. The other slopes to the S (Guggini, Seemate and Saenggini), follow a similar trend but with a slighter rotation. Dip angle values are higher to the bottom of the slopes and smoothly change upwards. Smaller values of dip are found on Randa, Hohebalme and Guggini. For the other two slopes in the S, the trend in the spatial distribution of the dip angles is similar.

A clear change in the behavior of the foliation is observed at the Blattbach River. To the N of the Blattbach River, there is no clear difference in orientations between bottom and top of the slopes. Medji can be considered as a clear bending point in the values of dip direction of foliation. As in the southern half of the area, to the S of Medji mean dip direction is towards the SW while to the N of Medji, foliation dips recurrently to the W-NNW. At the limit of the area (Sparru), a perceptible difference between bottom and top of the slope is observed. Bottom of the slope dips to the W, while to the top dips to the NW. Dip angle values have a clear change at the area of Medji. To the S
(Walkerschmatt), foliation dips gently with angles between 12° to 20° whereas to the N, dip angle raises up to more than 35 degrees to the NW of the area.

For this work, non-penetrative discontinuities parallel to the foliation in the area have been divided in two classes. A meso-scale fracture set that has been observed as well by other authors (Wagner, 1991; Willenberg, 2004), and a large scale set composed by ductile shear zones reactivated by brittle processes. The spatial distribution of these two sets cannot be clearly defined. They are present along the whole area, but spacing, persistence and other characteristics greatly vary along the area.

The other non-penetrative discontinuities were classified according to their persistence in meso and large scale discontinuities. They are basically brittle discontinuities that can be differentiated as joints, faults, shear zones and fracture zones. For the area joints are mainly meso-scale fractures; faults and fracture zones are large scale fractures. However, towards the N, especially after the Blattbach River, some joints can have persistence values up to hundred meters which should be classified as large scale fractures. However, for the present analysis fracture sets that were identified as meso-scale fractures in the southern part of the area were not later reclassified as large scale fractures in order to keep the coherence in the classification. In the case of fault and fracture sets, some faults and fracture zones can have persistence values of only few tens of meters, but they are simple specimens distributed along the whole study area and do not represent persistent changes in the characteristics of the fracture sets.

Four main meso-scale fracture sets have been found in the area, which have been described in detail above. They present clear spatial patterns and are persistent along the whole area. Spatial variation in their orientation values does not to follow a regular pattern. Some of the fracture sets present a slight clockwise rotation towards the N, but for some others a clear variation cannot be observed and it can even be considered random. This fact can be explained as a result of the high heterogeneity that fractures in general possess (Tran, 2007) and by the intrinsic errors and bias that databases have in general. However, a clear change in the relationships between sets can be observed at the Blattbach River. Frequency of fracture sets drastically changes, especially for sets J1 and J2. To the south, J2 is the most frequent set found in all datasets (spot and lineal mapping and DEM-based extraction data). To the N, contrarily, J1 becomes the most frequent set. As mentioned above, this change can have a connection with the rotation of the valley axis after the Blattbach River, which causes the migration of the exfoliation joints parallel to the valley axis from set J2 to set J1.

A variation in the number and distribution of fractures not related with the main sets is observed for the area. To the S, in the sub-domains Randa, Hohebalme and Guggini a higher number of
clusters exist and they are clearly observed within the field data and are slightly perceptible on the DEM-based data. That situation is due to the geometrical disposition of those fractures sets respect to the slope. Most of the above mentioned fracture sets are dipping into the slope and do not have a morphological expression on the current topography, in a similar manner to the conditions mentioned for the fractures parallel to the foliation. Therefore, they cannot be observed on the 3D shaded relief map and have a small representation in the DEM-based datasets. Other authors have reported up to 10 meso-scale fracture sets for those areas (Wagner, 1991; Willenberg, 2004). The increase in the complexity of the structural pattern for this part of the area can be related to two main factors: A first factor is the change on the lithological conditions respect to the rest of the area.

At the area of the current instability on Randa (Grossgufer), the presence of ortho- and para-gneisses has been reported by several authors (Bearth, 1964; Wagner, 1991; Sartori et al., 2003; Willenberg, 2004), whereas northwards, the area comprises only orthogneisses (the Randa Augengneiss). Paragneisses present a more irregular fracture network (Gischig et al., submitted).

The proximity to the confluence of the valley formed by the Bisgletscher glacier with the Matter Valley is considered a second factor of importance. Glaciations increase the load that rock slopes undergo causing gradual rock mass dilation (stress release), which causes, among other effects, propagation of the internal joint network (Ballantyne, 2007). A last factor only affecting the area of the current instability at Randa is the generation of secondary fractures product of intact rock bridges breakage caused by the strong deformation that the slope undergoes. Towards the northern limit of the study area, at Sparru, there is again a change on the lithology with sequences of mica-schists and quarzites to the bottom of the slope overlaid by ortho-gneisses in the middle part of the slope and para-gneisses to the top (above 2300 m asl). This lithological variation causes a more complex fracture network in a similar way that what have been mentioned for the area of Randa, Hohebalme and Guggini in the south.

Three main large scale fracture sets have been found in the area (besides an additional set, F1, parallel to foliation, mentioned before). F2 is composed basically by faults with gouge. F3 fractures are mainly fracture zones and it is similar in orientation to meso-scale fracture set J2. Faults and fracture zones are part of fracture set F4. Plane extraction from lineaments mapped during field work, image interpretation and DEM analysis show a similar trend. There is clear limit on the spatial disposition of the fracture sets at the Blattbach River. Large scale structures strongly control the current morphology to the north.
2.7.3 Implications for slope stability

An approach to slope stability for the area was carried out by a classification of the relationship between orientation of foliation and slope geometry. A classification proposed by Meentemeyer (2000) was used, which defines seven different classes. Cataclinal slopes have a higher probability of failure than the others classes (Cruden, 2003), being overdip slopes the sub-class with the highest probability of planar failure along the fractures parallel to foliation. Overdip slopes for the area, are mostly the south-faced slopes found to the S of Medji. According with Cruden (2003) toppling can as well be possible if further conditions are present (a second set of fractures dipping into the slope perpendicular to foliation).

Spatial variations of fracture characteristics as well as morphology of slopes have a strong influence on the stability. A higher level of activity is observed to the N of the Blattbach River, being morphologically expressed by a higher volume of debris deposits to the base of the slopes (Figures 2.23a and 2.23b). This limit coincides with the change of meso- and large-scale fracture characteristics. The analysis of the rock slope stability is presented in a companion paper.

2.8 Summary

A detail structural characterization for the western flank of the Matter Valley has been carried out for the present work. It has been focused in two of the main characteristics of fractures sets: orientation and length. Discontinuities have being divided on penetrative and non-penetrative. Penetrative discontinuities are represented by foliation in the area. Non-penetrative discontinuities comprise joints, faults, shear zones and fracture zones. Non-penetrative discontinuities have been sub-divided on meso and large scale discontinuities based on their length. Spatial variation on their orientation has been observed through a detailed analysis of data. GIS systems have been used to extract information from remote sensing imagery interpretation and DEM analysis. Reliability of these data has been tested by comparison with the field data using classic statistics, giving an acceptable matching.

Penetrative discontinuities have a uniform variation from S to N and from bottom to top of the slopes. Foliation has been divided in five domains based on its variation on orientation. Two sets of non-penetrative discontinuities are associated to foliation and present the same trend as foliation on their orientation. The rest of non-penetrative discontinuities present differences in their pattern between the northern and the southern parts of the area. Four main sets of meso-scale fractures and three sets of large scale fractures (additional to the sets parallel to foliation) have been found and described. A limit between southern and northern domains is located at the Blattbach River. The
variation on the fracture set pattern seems to be partially linked to the rotation of the valley axis
(15°, clockwise) what causes that exfoliation joints rotates in a similar manner. Variation in other
fracture characteristics and other fracture sets may be related with regional tectonics. Local
variations in the fracture pattern are linked with lithological changes.

Changes in the fracture network configuration have a relation with slope stability. To the N of
the Blattbach River, slopes have a higher level of rock fall activity than to the S due to the structural
predisposition.
2.9 References


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Figure 2.1. Situation Map of the study area.
Figure 2.2. Randa rockslides. a) Photograph of 1991 failure surface and deposits, current situation. b) Schematic representation of the situation before 1991 and the geometric disposition of the two failed masses of April and May 1991 (from Eberhardt et al., 2004). c) Cross section through the current unstable mass in the northern part of the 1991 scarp (from Gischig et al., 2009)
Figure 2.3: 2002 rockslide of Medji. a) Overview of the current situation b) Schematic cross section of the situation before the 2002 failure (From Rovina, 2002).
Figure 2.4. Foliation orientation and domains from field measurements.
Figure 2.5. Continuous grid surfaces representing spatial variation of dip (left) and dip direction (right) of foliation. Grids were built using disjunctive kriging.
Figure 2.6. Schematic representation of the relationships between topography and penetrative discontinuities (from Meentemeyer and Moody, 2000). Orthoclinic slopes are not shown.

Figure 2.7. Slope classification for the study area using the TOBIA categorical model.
Figure 2.8. Spatial distribution of the sampled locations within the study area.

Figure 2.9. Stereoplots of meso-scale discontinuities from field recognition for the study area (equal area, lower hemisphere). a) Results from scanlines southern domain (Zueger, 2007); b) Results from spot mapping southern domain; c) Results from scanlines northern domain; d) Results from spot mapping northern domain
Figure 2.10. Picture showing the influence of meso-scale fracture sets on the construction of landscape.

In the example, three sets control slope walls.

Figure 2.11. Changes observed on DEMs for the Matter Valley. a) View of orthophoto from the area. b) Photogrammetry DEM, c) LIDAR DEM, d) Comparison of profiles extracted from the DEMs (different scale for height and length).
Figure 2.12. Construction of a 3D shaded relief map. a) Example of 3D shaded relief map for the area of the Matter Valley, b) differences in representation between HSV and RGB color model schemes, c) Flowchart of construction using ArcGIS.

Figure 2.13. Principal source of errors a) Thick soil covers, b) Overhanging blocks on walls.
Figure 2.14. An example of a 3D shaded relief map showing fracture orientations in the northern part of the study area (Walkerschlatt). Color scheme has been inverted to help visualization. The stereoplot from DEM analysis to the right, and the stereoplot from field mapping to the left are shown (equal area, lower hemisphere projection). Smaller polygons represent areas selected for DEM analysis and the large polygon corresponds to limits of the sub-domain. Green dots represent field stations.
Figure 2.15. Areal correction for data extracted from a 3D color shaded relief map. a) Schematic representation showing the bias produced by slope angle between a projected surface and the real surface, b) Geometrical relationships between planes with similar area but different inclination angle.
Figure 2.16a. 3D shaded relief with stereoplots of meso-scale discontinuities to the south of the Blattbach River. Stereoplots obtained by DEM-based analysis to the left and by field recognition to the right. Large polygons represent limits of the sub-domains. Small polygons are sampled areas. Color of polygons is different for each sub-domain.
Figure 2.16b. 3D shaded relief with stereoplots of meso-scale discontinuities to the north of the Blattbach River. Stereoplots obtained by DEM-based analysis to the left and by field recognition to the right. Large polygons represent limits of the sub-domains. Small polygons are sampled areas. Color of polygons is different for each sub-domain.
Figure 2.17. Probability density function for dip direction of set 3 in the area of Walkerschmatt. Histogram represents the data obtained by DEM-based analysis. Lines represent five of the most representative fitted distributions for the sample.

Figure 2.18. Ranking of distributions using a) Kolgomorov-Smirnov test and b) Anderson-Darling test. The number on the y-axes represents the number of times a distribution was scored at the top three for all 78 samples.
Figure 2.19. Stereoplot showing large scale discontinuity orientations from field observation. Four main sets were found and described.

Figure 2.20. Fault Examples of large scale fracture sets. a) Ductile shear zone reactivated by brittle fractures in the area of Guggini (F1), b) fault with gouge and secondary structures (Riedels) lower part of Walkerschmatt (F2), c) fracture and fracture zones parallel to the valley axis (F3) and d) fracture zone in the area of the Guggigraben between Guggini and Seemate (F4).
Figure 2.21. Photograph taken in the area of Saenggini close to the Blattbach River showing the four main large scale fracture sets for the study area.

Figure 2.22. Stereoplot showing orientation for the lineaments identified from DEM analysis and photo-interpretation (stereonet equal area, lower hemisphere).
Figure 2.23a. Map of lineaments and their calculated orientations for the area to the S of the Blattbach River.
Figure 2.23b. Map of lineaments and their calculated orientations for the area to the N of the Blattbach River.
Chapter 3

Rockslide susceptibility as derived from integrated field and remote sensing investigations (Matter valley, Switzerland).

3.1 Introduction

Rock slope failures have always been considered as a threatening factor to the development and expansion of human settlements and vital networks and have caused substantial economical and life losses along historical times. For mountainous environments, rock avalanches and instantaneous rock and debris falls, slides and flows with long runouts are common types of slope failure as their morphology provide the potential kinetic energy requirements by having high relief and steep slopes (Glade & Crozier, 2004). More than 6% of the Swiss territory is prone to landslides (Noverraz & Bonnard, 1990 in Leroi, 2005). The highest concentration of the main landslides reported by Leroi is located in the south-western Swiss Alps. Based on the mobilized volume, rock slope instabilities can be divided in large (>10000 m³) and small instabilities.

Large instabilities can include different typologies such as rock debris flows (Sturzströme), translational slides, sagging (Sackungen) and deep seated deformations (Jarman, 2006). These types of instabilities start with relatively small displacements that can gradually develop to catastrophic stages. However, for some other cases, a new stability point can be reached with time. Large instabilities are structurally controlled (Agliardi et al., 2001, Brideau et al., 2009) and their origin has been postulated to be connected mainly to either or both, active faulting and post-glacial slope unloading. Usually, catastrophic failures of this type, when they occur close to human settlements, cause important damage on infrastructure and several life losses. Some of these events are counted among the most destructive natural events in mountainous areas in historical times (e.g.: Biasca, Switzerland in 1513; Elm, Switzerland in 1881; Vaiont, Italy in 1963; Huascarán, Peru in 1970).

Small instabilities are usually very rapid processes, with rockfalls being the most common type (Hantz et al., 2003). Volume frequency distribution of rockfalls has been studied by several authors (Hungr et al., 1999; Hantz et al., 2003; Malamud et al., 2004) who characterize the relationship between rockfall magnitude and frequency as following power law distributions. Hungr et al. (1999),
for two transportation corridors in southwestern British Columbia, observed that small rockfalls (<10000 m³) are the most recurrent phenomenon for natural and excavated rock slopes. For the same case, rockfall frequency was observed to be inversely proportional to its magnitude. Because of their high frequency, rockfalls are considered as an important factor on hazard and risk assessment in mountainous areas due especially to the increment on population and therefore, the expansion of human settlements (Leroi et al., 2005; Schuster & Highland, 2007). As a consequence, the identification of rockfall prone areas is of primary importance for landslide susceptibility assessment.

Landslide susceptibility is defined as the propensity of an area to generate landslides and can be used to predict the geographical location of future landslides. (Guzzetti et al., 2006). Susceptibility maps usually are the first stage in landslide hazard mapping, which is an essential part of quantitative landslide risk assessment (Erener et al., 2010). Several methods have been proposed to evaluate landslide susceptibility which can be grouped in two main types: qualitative and quantitative methods (Glade & Crozier, 2004). Geomorphological and heuristic are numbered among the qualitative methods. Quantitative methods are based on statistical classifications, process-based and numerical techniques.

In Switzerland, regional and local inventories at different locations have been carried out in the country in order to evaluate landslide susceptibility of rock slopes (e.g.: Baillifard et al., 2003; Loye et al., 2009). Other works have been focused on the detection or compilation of large instabilities (several cited on Noverraz & Bonnard, 1990). At 2005, 50% of the Swiss territory was already analyzed by landslide hazard maps at the scales from 1:25000 to 1:5000 (Lateltin et al., 2005). Besides this, several of the most threatening large rockslope instabilities have been studied in greater detail (e.g.: Randa, Preonzo, Val d’Inferno).

The Matter Valley between St. Niklaus and Randa is one of the deepest incisions in the Swiss Alps and an area that covers a very high density of natural hazards, including snow avalanches, floods and various types of landslides. In this area several maps about rockfall distributions (e.g.: King et al., 2001), and landslide hazards (e.g.: Roulier & Jaboyedoff, 1998) have been created in the past, but these studies have mainly focused on the development and testing of methodologies, rather than on the causes of rock slope failures. In this project we aim to improve our understanding of the predisposition factors and mechanisms leading to rock slope failures. While chapter 1 of this PhD describes in great detail all structural elements critical for small and large scale rock slope instabilities, this second chapter describes the spatial and temporal distributions and the mechanisms of past and ongoing large scale slope instabilities and failures. A special focus of the investigations presented in this chapter is on the role of large scale fractures (faults) controlling
large scale rock slope instabilities. In chapter 3 spatial distributions of small scale instabilities will be presented.

For the present work, a large scale inventory and map of landslide phenomena (instabilities, release areas, deposits, stable slopes) was created based on intensive field work, interpretation of airborne- and terrestrial-based remote sensing imagery and digital elevation model (DEM) analyses (section 3). Rock slope instabilities in the area were classified based on their volume in small (<10000 m³) and large events (Hungr et al., 1999; Fischer, 2009). Additional work was carried out in four of the current large instabilities detected for a better understanding of their mechanisms, by detailed field recognition, photo-interpretation, monitoring with differential GPS, geodetic networks, and ground- and satellite-based InSAR techniques (section 4). An empirical interpretation of the resultant inventory was carried out in order to assess the spatial distributions and causes of larger landslides in the study area (section 5). Geographic Information Systems (GIS) were used to integrate all different information in order to facilitate the analysis and presentation of results.

### 3.2 Landslide inventory generation

According to Malamud et al. (2004), landslide inventories can be divided in two classes. a) Landslide event inventories associated with a trigger event, and b) historical landslide inventories which are the sum of all landslides occurred over time in a region. For both cases, recognition of geomorphological features is the main criterion for landslide identification. Landslides inventories, depending on the scale, can represent landslides with more or less detail. Usually inventory detail is selected according with the project scale. Singhroy (2009), states that projects can be classified based on their scale on global, national (1:1000000), regional (1:100000), medium (1:25000-50000) and local (>1:15000). Local and medium scale inventories can show different landslides components as scarps or deposits, whereas small scale inventories usually show only location of landslides represented by points (Hervás & Bobrowsky, 2009).

Geomorphological analysis for landslide susceptibility is the simplest of the qualitative methods (Aleotti & Chowdhury, 1999). Assessment is carried out by an earth scientist based on his/her expertise and supported by detailed photo-interpretation and field work. A strong component of subjectivity in the selection of the environment factors is implied. However, it allows a rapid assessment of stability and it is feasible to be applied at any scale. Heuristic methods reduce the subjectivity observed on the above mentioned methods by adding to the analysis weighted maps of factors associated with landslide occurrence (Hervás & Bobrowsky, 2009). Weights are assigned
based on personal criteria of the project responsible. Subjectivity can be further reduced applying semi-quantitative models based on multi-criteria evaluation (Barredo et al., 2000).

Quantitative methods tend to be based on objective criteria and be created with the purpose of being reproducible for other cases with similar data sets. Statistical approaches are among the most common quantitative methods (Hervás & Bobrowsky, 2009). Statistical models can be sub-divided on bivariate and multivariate analysis models. For bivariate methods each factor map is combined with landslide inventories and is weighted independently. Multivariate analyses differs from bivariate analyses on that it determines the relative contribution of each slope instability map factor to the final susceptibility maps via different techniques (e.g.: discriminant analysis, multiple regression, artificial neural networks). Physically based models are built based on physical laws influencing slope instability. These types of methods are used in order to analyze a specific type of landslide or to investigate the effect of a particular trigger (Hervás & Bobrowsky, 2009). A complete data set of material properties and triggers are needed, which makes its implementation costly and time consuming.

As mentioned before, remote sensing imagery and DEM analysis have become important tools for landslide susceptibility and landslide detection in recent years. Airborne-based optical imagery is usually used to detect morphological features that represent slope movement (e.g.: scarps, counter scarps, cracks) on large instabilities allowing the production of phenomenological maps. For the case of smaller instabilities, its main function is their detection and contouring. Stereoscopic analysis of aerial photographs is a standard technique used for these purposes. For the case of aerial digital photographs, computer-based analysis can be done with the creation of anaglyph views, 3D visualizations (orthoimages draped on DEMs extracted from the same set of images), and stereo viewing of epipolar images. The recent advance on technology has greatly improved the spatial resolution of satellite-based optical images. Several high resolution satellites have been launched during the last decade (e.g.: IKONOS, QuickBird), allowing to cover wider areas with a less number of images making the work easier and more affordable. For some of the high resolution satellites, stereoscopic analysis is also possible with similar tools as the previously mentioned for aerial digital photographs (Nichol et al., 2006). High resolution imagery is usually thought only useful for local detailed studies, but Liu et al. (2004) have pointed out that even for regional studies local topographic detail is needed in order to detect landslides in steep minor valleys. Havenith et al. (2006) found that for an inventory of landslides generated from a 15 m resolution ASTER set of images and its accompanied 30 m DEM, landslides smaller than 7000 m² could not be fully detected and mapped.
Airborne-based LIDAR surveying is another remote sensing technique that has been used in the last years in order to create high resolution digital elevation models (HR-DEMs). HR-DEMs are important source of data for landslide inventory creation. Several examples of inventories for different types of landslide phenomena are found in literature. They range from shallow landslides triggered by heavy rainfall events to deep seated instabilities (Ardizzone *et al.*, 2007; Van Den Eeckhaut *et al.*, 2007; Schulz, 2007; Kasai *et al.*, 2009). LIDAR based DEMs allow the removal of vegetation which produces an enhancement of the morphological features and, therefore, for some cases, an improvement in the identification of slope instabilities. Schulz (2007) identified four times more landslides for the area surrounding Seattle and Haugerud *et al.* (2003) two times more deep seated landslides in the Puget lowland area close to Washington using LIDAR HR-DEM than with traditional aerial photo-interpretation.

Active satellite imagery has been incorporated as a tool for landslide detection and monitoring in the recent years. The multi-incidence angle, stereo and high resolution capabilities that radar satellites provide have been proved useful for landslide inventory maps (Singh *et al.*, 2005). Some authors have shown that high resolution stereo synthetic aperture radar (SAR) and optical images, combined with topographic and geological information have assisted in the production of landslide inventory maps (Singhroy *et al.*, 1998; Singhroy & Mattar, 2000 in Singhroy, 2009; Catani *et al.*, 2005). Differential interferometric SAR techniques have been used for landslide displacement rates estimation, because they can generate wide-area maps of ground surface deformations with precision in the order of millimeters (Colesanti & Wasowski, 2006). However, coherence loss (especially in vegetated areas) and atmospheric effects are the most important constraint factors to be taken in account before the application of these techniques. Displacements are measured on the line of sight direction given only a partial view of the slope instability velocity vector (Singhroy, 2009).

In recent years ground based remote sensing techniques for landslide analysis have had an important development. Optical and radar based techniques (including LIDAR) have developed under the same principles as the airborne- and satellite-based afore mentioned (but with different implementation) and have demonstrated to be useful for site monitoring, and structural and morphological analysis of large rock slope instabilities (Sturzenegger & Stead, 2007; Gischig *et al.*, 2009).
3.3 Landslide inventory of the Matter Valley

For the case of the western flank of the Matter Valley between St. Niklaus and Randa, a historical inventory of slope instabilities was created based on field recognition; photo-interpretation of two sets of aerial digital photographs (2001 and 2005, provided by SWISSTOPO), DEM analysis, and analysis of photographs acquired on field (see also chapter 1 for data sets and processing methods).

The inventory includes rock slope instabilities and deposits. In total seven different classes of slope instabilities and eight classes of deposits were used in order to create a landslide map of the study area (Figure 3.1) that was later used to investigate the scientific questions presented in section 1.

3.3.1 Deposit Units

The transported rock deposits found in the area have been classified according to the deposition type and process. Four main types of deposits were found: rectilinear talus slopes, debris cones, alluvial fans and glacial moraines. Besides, special deposits were included in separated classes (i.e.: composite cones, debris on slope, deposits of undefined origin). Mapping and contouring of the deposits were carried out by photo interpretation and DEM analysis. For the image analysis, an orthophoto mosaic created with aerial photographs acquired in 2005 provided by SWISSTOPO was used. A 3D shaded relief map (Jaboyedoff et al., 2004) was constructed from a high resolution digital elevation model (HR-DEM) of 2.5 m pixel resolution. The 3D shaded relief map gives the opportunity to easily differentiate and delineate deposit shapes based on the change of color and texture (Figure 3.2a and 3.2b).

3.3.1.1 Rectilinear talus slopes

For the present work, all debris deposits that do not show lateral variation of their geometry have been considered rectilinear talus slopes. They are usually products of discrete rockfalls on un-bisected slopes of monolithic composition (Luckman, 2007). For the case of the Matter Valley it was considered that the process started after the last glacial maximum (LGM) around 20000 years B. P. (Kelly, 2004).

Rectilinear talus slopes are found in the southern part of the study area, on the slopes of Hohebalme and Guggini, and in the northern part in the slopes of Walkerschmatt and Sparru (Figure 2.1 of Chapter 2). In the central part of the study area, no rectilinear talus slopes exist. An example of a rectilinear talus slope deposit is shown on Figure 3.2d. Talus slopes lay beneath steep rock walls with high relief with slope angles around 70°. They have an average horizontal length...
that can vary between 50 m and 200 m (Figure 3.3c). In terms of their geometry, the top of the deposits has a concave shape with a slope dip angle of around 40° to the uppermost part, followed downwards by a long regular steep slope of around 35° that prolongs until the toe of the deposit where flattens to slope dip values close to horizontal (Figure 3.2d). The described geometry is similar with what has been reported in previous works (Kotarba & Strömquist, 1984; Curry & Morris, 2004; Luckman, 2007). Most of the areas covered by such deposits are forested slopes which indicate a low state of activity nowadays. However, some non-vegetated patches in the deposits (Figure 3.1a) can be considered as evidence of recent activity. The areas occupied by this type of deposit are often intercalated with debris cones.

3.3.1.2 Debris cones

Debris cones are defined as debris deposits formed mainly by two different type of processes: 1) recurrent small rockfalls events, which deposition is controlled by gullies that dissect the source rock walls, and 2) large size single rockfall or rock slide events, that are not transformed into rock avalanches. The latter usually start on an un-bisected rock wall. In the study area, the debris cone formed by the Randa rockslides in 1991 is a good example. For the former, besides small rockfalls, debris can also be delivered by other gravitational processes such as snow avalanches or debris flows that change the shape of the lower part of the cone (Luckman, 2007).

Debris cones in the study area are widely spread, mainly located to the front walls of the valley flanks. A minor group of cones are on heights above the valley bottom and have been formed by single rockfall/rock slide events (e.g.: the deposit to the east of Topalihuette). In general the magnitude of the deposits (quantified by their area) increases towards the north. There is an evident change on the magnitude of the deposits at the Blattbach River (Figure 2.1 of Chapter 2). Debris cones to the north of the Blattbach River have, in some cases, twice the magnitude of those located to the south. Differences in magnitude are related with an increment in the rockfall activity to the north and a more abundant presence of relict and recent large size rockslide events. Because of their larger magnitude, there are several coalescent cones to the north of the Blattbach River. The length of debris cones fluctuates between 50 m and almost 600 m, being for most of the cases of values between 100 m and 200 m (Figure 3.3a).

Debris cones are often located at the exit of gullies excavated on steep rock walls (Figures 3.2a and 3.2c). Slope angle of the gullies are around 55°. Cones have a concave shape to the top with a maximum angle of 40° followed by a steep straight slope of around 35°. Even if in appearance debris cones have smooth surfaces, they present several more fluctuations along their downward profile than the talus slopes deposits represented by higher variations of slope dip angle values along the
longitudinal profiles (blue lines on the Figures 3.2c and 3.2d). These variations are caused by recent rockfall activity and other transport events which increase the slope surface roughness. Debris cones possess a concave shape at the toe of the deposit. Sometimes, reshape of the top of the cone may be produced by fluvial incision. Starting time for the debris cone development in the area is similar to what is considered for the rectilinear talus slopes (end of the LGM).

3.3.1.3 Debris flow cones

*Debris flow cones* are fan-shaped deposits of water-transported (“alluvial”) material supplied by areas found far in the upper part of the rock slopes (Bollschweiler *et al.*, 2008). The material reaches its current location transported by high energy geomorphic processes mainly from debris flows. Several studies have been carried out to characterize this type of deposits in eastern flank of the Matter Valley (Rebetz *et al.*, 1997; Bollschweiler *et al.*, 2008). The deposits are formed by several successive events, some of which include channel incision, which gives them a more undulating morphology than the previous type of deposits (Figure 3.2e). Several abandoned channels and lobes are found on the deposits (Figure 3.2b). Locally older deposits have been eroded by more recent events of the same type.

Debris flow cones are located at the outlets of the main lateral streams in the area. Their magnitude (area) is slightly larger in the northern part of the study area (to the north of the Blattbach River). Slope angle is regular along the deposits with a slight decrease on the slope angle values to the toe of the deposits (Figure 3.2e). Average slope angle values vary from one deposit to the other, ranging on values from 10 to 30 degrees. Length of the fans varies from 200 m to 580 m being their mean length about 300 m (Figure 3.3b). Recent debris flow activity on the cones can lead to headwater gullies producing an irregular pattern of debris flow sediments.

Small rockfalls at the flanks of the main lateral streams are one of the most important sources of sediments for debris flows in the area. They produce the source material that once is accumulated on the stream beds is remobilized downwards by debris flows triggered by an increase on their water flow and/or a bigger rockfall event upwards. Debris channels on this work do not include all debris flow transit areas but only the places where some material have been observed to be accumulated.

3.3.1.4 Moraines

Moraine deposits found in the area are probably corresponding to the Clavadel/Senders (16-14.7 ka) or the Egesen glacial stadials (14-12.9 ka). Equilibrium line altitude (ELA) depression of the Clavadel/Senders stadial respect to the Little Ice age (LIA) has been reported to be of 400-500 m
with lower values in the more sheltered positions (Ivy-Ochs et al., 2008). In the case of the Egesen stadial, ELA depression in the western and northern Alps may be up to 400 m (Ivy-Ochs et al., 2008). Additional work is needed for a more precise definition. No clear evidences of moraines were found within the study area in the Matter Valley but presence of lateral and terminal moraines were found in some of the lateral hanging valleys that bisect the flanks of the Matter Vispa River (Figure 3.4). They are not deposits product of rock slope failures but were mapped during the present work in order to differentiate them from the landslide-related deposits. Morainic deposits can be differentiated by DEM analysis because they present a distinctive and contrasting signature respect to the landslide related deposits (smooth surfaces). The deposits have no regular shapes and are not fully preserved as they have been eroded by other processes such as debris flows and slides (Figure 3.4b).

3.3.1.5 Composite deposits

Composite deposits correspond to deposits formed by at least two different types of denudation processes and no clear control of a single process in the development of the deposits is evident. In the study area, there is one deposit that has been classified as composite deposit which is located beneath the Medji rockslide (Figure 3.5). The morphology of this deposit is composed of three main lobes and has been reworked by human activity what makes difficult to observe its original characteristics. The southern part of the deposit has a clear colluvial component with a high slope angle and a smoother surface (Figure 3.5d), similar to the debris cones deposits (Figure 3.2c), but with a smaller average slope angle (~25°). The absence of a clear source area for the debris in the surroundings implies a long transport from areas far upwards. The northern two lobes have a signature of debris flow cone deposit, but with a more elongated shape and a lower average slope angle. In the upper part of the northern area of the deposit, average slope angles are higher than in the lower part (Figure 3.5c) with slope angle values similar to what is observed on the southern part of the deposit (~25°) whereas on the lower part more gentle values are found (mean value ~12.5°). The far north lobe is still active as new material is transported from the upper slope by one of the lateral streams of the study area. Evidences of recent debris flow events are observable on this lobe too (abandon channels, levees, lobes). The magnitude of this deposit is clearly larger than any other debris cone or debris flow deposit in the study area (area~350000 m², maximum length~850 m). Its magnitude is only comparable to the magnitude of the deposit produced after the 1991 Randa rockslide events (area~ 580000 m², length~1050 m).
3.3.1.6 Debris on slope

This type of deposit is a minor class for the area. They are comparatively small volumes respect to other deposit units (shallow deposits). These deposits are located on several slope shoulders on stepped slopes (Figure 3.6), are formed by several types of processes such as snow avalanches or small size rockfalls, and correspond to rock debris that cannot reach the slope bottom because of the stepped morphology and reduced volumes. They are more abundant in the southern part of the area (to the south of the Blattbach River), where walls parallel to the axis of the main valley or side walls of lateral streams present a stepped morphology. The deposits are located preferentially on the northern side walls of the secondary lateral streams, because often they present a stepped morphology.

3.3.1.7 Undefined deposits

These are deposits that have been observed only on the aerial photographs and/or the 3D relief map with no field recognition information, which make difficult their classification into the other deposit units. This unit is formed by deposits possibly created by similar processes as the deposits from the other units but with no enough evidence to be assigned to one of the other classes.

3.3.2 Slope instability units

Rock slope instabilities were classified according to two different criteria: magnitude and state of activity. Several classes were defined based on these two criteria. An assessment of the magnitude of slope instabilities was carried out based on the area occupied by the phenomena. For small instabilities (less than 10000 m³), no additional classification was carried out. Large instabilities were classified based on their state of activity as past or current. Past instabilities correspond to release areas were features of former failures could be observed or inferred from DEM analysis using a 3D relief map (Jaboyedoff et al., 2004). They were divided according with their age in lateglacial and post-glacial. Current instabilities comprise areas on slopes that show clear signs of movement without failure. They were divided based on the depth of their failure surface in shallow, intermediate and deep instabilities. Swiss federal regulations classified shallow landslides as the landslides events with a depth of the slip surface smaller than 2 m, intermediate landslides are landslides with depths ranging from 2 m to 10 m, and deep landslides all events with depths larger than 10 m (Lateltin et al., 2005). Landslide depths accompanied with landslide velocities are used to assess the potential danger of slides. This classification is intended to contribute on the generation of hazard maps for land use planning and do not discriminate between soil cover slips (superficial phenomena) and bedrock failures (larger range of depths). For this work, it was considered that those classes do not fit with the phenomena observed in the Matter Valley, where rock slope
instabilities range from very superficial (shallow rockfalls) to deep seated slope gravitational deformations (e.g. slope instability at Graechen). In the area, shallow instabilities are considered rock slope instabilities that have a slip surface up to 10 m. They are mainly rockfalls from steep rock walls. Intermediate instabilities are considered those instabilities where a clear basal limit exists, even if there is no clear surface delimiting the mass. Depth of their basal limit can vary from 10 m to around 150 m for the area, like in the case of Randa (Gischig et al., 2009). Deep instabilities are considered rock instabilities that fit with the definition presented on Agliardi et al. (2001): “slope movements occurring on high relief-energy hillslopes, with size comparable to the whole slope, and with displacements relatively small in comparison to the slope itself”. Several of the current and large instabilities were studied in more detail and are extensively described in further sections of this chapter.

Finally, slopes not affected by large instabilities were considered stable. However, the occurrence of small size instabilities was used as a parameter to qualify stability of the overall slopes: stable slopes with rockfalls and without rockfalls. Areas of difficult access that do not present clear features of instability or are covered by vegetation on aerial photographs were classified as undefined.

3.3.2.1 Release areas of lateglacial failures

This unit is formed by rock slope failures occurred during the Lateglacial period (Alpines Spaetglazial) after the Late Glacial Maximum (LGM) occurred circa 20000 yr BP (Kelly et al., 2004). The Lateglacial comprises the period where glaciers re-advanced several times (stadials) until glacier reached the dimensions of their maximum extent during Holocene (Ivy-Ochs et al., 2008). The failures can have occurred on paraglacial conditions after the LGM or during interstadial periods. Several authors (Ballantyne, 2002; Jarman, 2006; Curry & Morris, 2004) have postulated that major failures on formerly glaciated valleys occurred during a brief period after deglaciation. Lateglacial failures were primarily identified by abrupt changes of the rock slope geometry (e.g. sudden changes in slope angle) and by evidences of non-glaciated surfaces on rock walls. Most of these areas have a difficult access and their identification has been done mostly by photo interpretation and DEM analysis. Lateglacial failure surfaces do not have a distinct associated debris deposit, probably because they were washed out by later glacier readvance during one of the subsequent stadials or covered by more recent sediments on the valley floor. For the present work, no information of subsurface investigations on the valley bottom for the area was available. No other authors have reported data about thickness or shape of the valley fillings. Lack of associated deposits is evident for the failure surfaces found to the south of the Blattbach River, however other morphological features can be clearly observed. In the case of the deposits to the north, it is not differentiable if the
deposits from Lateglacial failures still remain because of thick layers of more recent minor size rockfall deposits do not allow to isolate deposits of relic large events.

Surface area values of the relic failures vary between 2500 m² to 190000 m² (Table 3.1). Surface area is defined as the exposed area of a solid body. In this case, the exposed area corresponds to the release area of the rock slope failure event. Surface area values were calculated using standard tools on ArcGIS from a 2.5 DEM. Surface area was chosen to define the magnitude of the relic failure surfaces due to the steep morphology of the scars left by the rock failure events. In some cases, planimetric area values were up to eight times smaller. Nine of a total of fifteen relic failures were found in the southern part of the area (south of the Blattbach River). The larger failure areas are as well found in the southern part, where most of the areas are above 25000 m² (Table 3.1). To the north, most of the values are between 12000 m² and 35000 m², with the exception of the failure surface located on the southern rock wall of Walkerschmatt, which is the second largest in the whole area (~185000 m²). There is no a good estimation about the depth of the failures as no slope reconstruction was carried out during the present work, but it has been assumed that thickness of the mobilized masses were in the order of several tens of meters.

3.3.2.2 Release areas of post-glacial failures

Rock slope failures that have occurred during the Holocene (after glacier stabilization) on the rock slopes in the area have been classified as recent failures. In general they occupy smaller areas than the relict failures and are widely spread within the study area. They are distinguishable by a marked change in the color of the rock exposure (Figure 3.7). The rock faces have a lighter color than glaciated rock surfaces and sometimes dark stains, produced by water circulation, on the rock face. For most of the cases, a structural control of the failure surface is evident. Only the larger recent failure surfaces are shown on the map presented in Figure 3.1. Several of the small failure surfaces cannot be mapped at the presented scale. However, they were detected and mapped on more detailed elevation models carried out using ground based digital photogrammetry methods on the lateral walls of the Matter Valley (an example can be seen on Figure 3.7). Several failure areas are not observed on orthophotos or aerial-based DEMs because they are occluded to the sensors by remnant overhanging blocks on top of them. 3D visualizations of the ground based elevation models (Figure 3.7b) were useful for a complete overview of small rockfall release areas.

A larger amount of failure surfaces at all scales were observed on slopes located to the north of the Blattbach River. As an example, 48 mapped recent release surfaces shown in Figure 3.1 are found to the south of the Blattbach River whereas 77 recent release surfaces have been mapped to the north. The latter suggests a higher level of activity on recent times in the northern part. These
failures are gravitationally driven processes and they are produced in places with a favorable morphology and structural predisposition.

Most of the deposits related to recent smaller release surfaces have not been individually identified because of their small volumes that deposited over other deposits (especially debris cones). In few cases when failures have occurred on stepped slopes, part or all of the material released is deposited on the slope shoulders which correspond to the “Debris on slope” unit. Several of the larger postglacial failures can be associated with deposits to the toe of the rock walls, especially to the south of the Blattbach River where the volume produced by small size rockfall events is smaller. To the north, because of the larger volume of debris produced by small size events, identification of large failure deposits is not possible.

3.3.2.3 Current instabilities

In the area several regions have been identified as presenting signs of ongoing deformation. The current instabilities have been classified in deep, intermediate and shallow based on an estimation of the depth of movements (see map, figure 3.1). The most prominent current instabilities are located at Randa, Medji, Walkerschmatt, Sibulbodme and Topali. These areas have been the subject of detailed field mapping and monitoring in order to have a better understanding of their geometrical and structural characteristics and failure mechanisms. The most recent investigations carried out at Randa are described in Gischig et al. (2009, 2010) and not recalled herein.

3.4 Detailed analysis of selected rock slope instabilities

3.4.1 Displacement monitoring

An important part for the understanding of the kinematics of rock slope instabilities is based on displacement monitoring. For the area of the Matter Valley, several monitoring systems have been applied in the selected unstable areas. A summary of the available data from monitoring in the study area follows:

A Differential Global Positioning System (D-GPS) network was implemented on four of the largest slope instabilities (Walkerschmatt, Sibulbodme, Topali and Sparru). The system was designed and implemented in collaboration with the group of Geodetic Metrology and Engineering Geodesy, ETH Zurich during summer 2008. The monitoring was carried out in order to obtain an overview of the rate and direction of movement for all selected areas and to get information about internal kinematics of the rock slope instabilities. The network was composed of three base stations located in the valley bottom and thirty-five stations unevenly distributed on the selected
instabilities (Figure 3.8). All points were surveyed during two campaigns carried out within an interval of one year (summer 2008 – summer 2009). Static, rapid-static and kinematic methods were used for the surveying (for a detailed description of the methods see Gili et al., 2000). The method selection was based on the accuracy required according to the expected displacement rates (few cm per year) and time efficiency. Measurements were carried out with 4 to 8 Leica system 1200 devices and a frequency of 0.2 Hz (static measurements) to 1 Hz (kinematic measurements). Accuracy of the results is in average 5 mm for planimetric values and 1.3 cm for 3D values. Results and analysis are shown in detail for each monitoring area in further sections of this chapter.

<table>
<thead>
<tr>
<th>Slope Name</th>
<th>Failure code</th>
<th>Planimetric Area (m²)</th>
<th>Surface Area (m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hohebalme</td>
<td>HH1</td>
<td>7184.89</td>
<td>23485.2</td>
</tr>
<tr>
<td></td>
<td>HH2</td>
<td>644.16</td>
<td>2427.31</td>
</tr>
<tr>
<td></td>
<td>HH3</td>
<td>3361.76</td>
<td>28845.58</td>
</tr>
<tr>
<td>Guggini</td>
<td>GG1</td>
<td>54298.08</td>
<td>116939.84</td>
</tr>
<tr>
<td>Seemate</td>
<td>SM1</td>
<td>98441.42</td>
<td>191275.78</td>
</tr>
<tr>
<td></td>
<td>SM2</td>
<td>20856.51</td>
<td>27726.16</td>
</tr>
<tr>
<td>Saenggini</td>
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</tr>
<tr>
<td></td>
<td>SG2</td>
<td>19104.7</td>
<td>37766.1</td>
</tr>
<tr>
<td></td>
<td>SG3</td>
<td>19385.8</td>
<td>37766.1</td>
</tr>
<tr>
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<td>96944.88</td>
<td>184052.47</td>
</tr>
<tr>
<td></td>
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<td>34352.65</td>
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<tr>
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<td></td>
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<td>8496.86</td>
<td>12512.06</td>
</tr>
</tbody>
</table>

Table 3.1: Planimetric area and surface area values of the Lateglacial rock slope failure surfaces shown in the Figure 3.1.

Satellite Based Differential Interferometry Synthetic Aperture Radar (SB- DInSAR) data was used for regional verifications and local site investigations at the selected instabilities. The data was acquired within the framework of the SLAM project (Service for Landslides Monitoring), a project
founded by the European Space Agency (ESA). Data was provided by the Swiss Federal Office for the Environment (BAFU) and comprises the results from the application of Interferometric Point Target Analysis (IPTA) to a series of 30 snow-free SAR images acquired between 1995 and 2000 in ascending mode to the Matter Valley. IPTA is “a method to exploit the temporal and spatial characteristics of interferometric signatures collected from point targets to accurately map surface average deformation rates, deformation histories, terrain heights, and relative atmospheric path delays” (Manunta et al., 2003). The images were reprocessed at local scale for the sites of Medji and Randa where several additional targets were collected giving a more accurate image of the ongoing displacements of the instable rock masses. For the rest of the area only regional scale data was available with a smaller number of targets and a rougher processing. Analysis of its results in combination with other sources is further shown in subsequent sections.

Ground based Interferometry Synthetic Aperture Radar (GB-InSAR) data was used for the instabilities at Medji. The data was provided by LISA Lab and produced within the framework of a BAFU project funded by ASTRA (Bundesamt fuer Strassen). The data was collected during two campaigns carried out on September and November 2005 with an interval of approximately 60 days. The system parameters for the area of Medji are shown in Table 3.2.

Geodetic measurements of several instabilities have been carried out in the area in different periods. The measurements include 1D and 3D displacements at Medji and Walkerschmatt as measured between the years of 1995 and 2009. For the area of Medji, several 3D monitoring points were installed few months before the rockslide occurred in November 2002 and most of the stations were located within the unstable mass. After the event, new stations were installed to monitor already detached blocks that can potentially threat people and infrastructure at the valley bottom. For Walkerschmatt, 3 stations have been monitored not periodically since 1999. Two of them are located out of the instable mass. Results, in combination with other data for the selected rock slope instabilities, are presented in further sections.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Synthetic aperture length (cm)</td>
<td>270</td>
</tr>
<tr>
<td>Bandwith (MHz)</td>
<td>80</td>
</tr>
<tr>
<td>Antennas beam-width (˚)</td>
<td>30</td>
</tr>
<tr>
<td>Antennas inclination (above horizon) (˚)</td>
<td>20</td>
</tr>
<tr>
<td>Distance to the surveyed area (m)</td>
<td>1000-1900</td>
</tr>
<tr>
<td>Width of the surveyed area (m)</td>
<td>900</td>
</tr>
</tbody>
</table>

*Table 3.2. Parameters utilized for the GB-InSAR surveys carried out in the area of Medji.*
3.4.2 Walkerschmatt

The area of the instability at Walkerschmatt (Figure 3.9) is located at a height ranging from 1855 m asl to 2130 m asl on a ridge with a gentle slope face on top and bounded by two steep cliffs facing to the south and east. The instability is lying on the crown of a Lateglacial failure surface to its south (see map Figure 3.1). Slightly or not weathered Randa augengneiss is the rock type for the whole rock mass. Shear zones occurred parallel to foliation which are formed by a layer of, sometimes heavily weathered, mylonite of different thickness. Foliation orientation is similar to the trend for the area (240/21). Another fracture set (small- and meso- scale) is parallel to foliation too. Within the unstable mass there are two predominant fracture sets which correspond to similar sets as J1 (132/85) and J4 (024/83) presented in Chapter 1. Set J2 (067/72) is present with a lower density (Figure 3.9a). A fault (036/41) from set F1 (see chapter 1) is proposed as a possible basal limit of the instability. A series of faults corresponding to this set are visible on the eastern face of the rock mass.

Ambient vibration measurements were carried out in August 2009 at Walkerschmatt. The experiment was executed in a similar way as it was previously carried out at the current instability of Randa (a detailed description of the methodology can be found in Burjánek et al., 2010). Three arrays with a total of 28 positions were surveyed with 12 three-component velocity sensors. A station installed close to a small hut at the top of the instability, in an area considered as stable, was used as the reference station. The position of each node was collected using differential-GPS. Figure 3.9b shows site-to-reference spectral ratios calculated comparing station response within the instability to the reference station in the stable area near the hut. The results show clear polarizations of spectral ratios for a characteristic frequency ~1.5 Hz for most stations (Figure 3.9b). The wave field is dominated by normal-mode vibration of rock blocks or structural compartments exhibiting similar behavior. Based on the results it is possible to differentiate three compartments of the moving mass.

Compartment I is located at the southern border of the rock slope instability, bounded by a steep cliff striking WNW-ESE (azimuth ~120°) with shallow debris deposited at its bottom (Figure 3.9b). Only one station was located in that area during the ambient vibration surveys; it registered noise polarization in the highest frequency of about 30 Hz. The direction of its maximum noise polarization is about 170° and perpendicular to the mentioned rock cliff at the station location. Measurements of the opening of displaced fractures in this compartment show the highest values of the whole site. Moreover, the open fractures have a preferential orientation of ~020/80 (WNW-ESE), perpendicular to the maximum noise polarization. Fractures from this set (set J4, see chapter 1)
constitute the northern boundary of the compartment. A series of widely open fractures from sets J4 and J1 can be observed into the compartment.

Compartment II corresponds to the lower part of the instability towards its eastern border. Its upper boundary is controlled by a set of J1 fractures and the lower boundary corresponds to a steep cliff striking NE-SW. The compartment shows a stepped surface morphology, where the steps are limited by meso-scale fractures from set J1. According with the results from the ambient vibration tests, directions of maximum noise polarization at 1.5-3 Hz strike perpendicular to the fracture set J1 (azimuth ~120°) and show amplifications of horizontal velocity with respect to the stable reference station ranging between 2 (west) to 8 (east).

Compartment III corresponds to the upper part of the unstable area. A clear upper boundary is not defined. The lower boundary is defined by a series of widely open J1 fractures. There is no clear evidence of a well developed failure surface daylighting at the back of the instability. According to the results from the ambient vibration measurements there is not a clear pattern for noise polarization directions within this compartment. However, near of its lower boundary, maximum noise polarization follows a similar trend as in compartment II. Towards, the E, polarization is more difficult to define. The main fracture sets are still J1 and J4 (Figure 3.9b) but with apertures of the fractures drastically reduced.

Results from the differential-GPS monitoring for the area do not show conclusive results (Figure 3.9c). Five stations were installed in the area and surveyed two times with an interval of approximately one year. 4 of the 5 stations registered displacements lower than the accuracy of the method calculated for the area (1.0 cm). Only one station (404, displacement 1.28 cm) registered a displacement larger than the accuracy which shows a NE (N25°) direction of displacement. Cumulative displacements of rock blocks calculated based on measurement of open joints bounding displaced blocks, collected during field recognition shows N-S to NW-SE trend (yellow arrows in Figure 3.9c). Opening of joints corresponding to set J1 are larger, and range between 0.3 to 2 m, for J4 fractures opening varies between 0.35 and 0.8 m. Opening on J2 fractures is no common, being around 0.5 to 0.8 m.

1-D geodetic monitoring of three stations has been carried out for a period of 14 years (starting at June 1996) with a varying frequency ranging from 2 weeks to almost two years. Only one of the three stations is located inside of the unstable mass (station 31, represented by a red square in Figure 3.9a). The displacement was measured in a direction striking ~100° (close to perpendicular to the main direction of movement detected by D-GPS). Maximum cumulative displacement was observed on July 2002 (0.003 m). Displacements measured on this station are considered unreliable because measured values are below the accuracy of the instrument (0.002 m + 3 ppm).
Kinematic analysis was performed in order to assess possible movement and failure mechanisms for the three steep edges of the rock slope instability (Figure 3.9c). A slope oriented 115/82 which corresponds to the average orientation of the face of the slope parallel to the Matter Valley axis located to the east, another slope orientated 190/63 which corresponds to the average orientation of the southern edge of the site, and a slope orientated 060/80 corresponding to a NE facing slope similar to the northern boundary of the ridge. For all cases three fracture sets were analyzed (J1, J2 and J4) plus the foliation parallel set using 34 orientations collected during field recognition. A conservative basic friction angle of 26° (for smooth surfaces) was chosen from the literature (Priest, 1993) for materials of similar characteristics to the Randa Augengneiss.

For the first slope geometry (left column in Figure 3.9c), planar and wedge failure might occur along the set J1 and J2 and the intersection J1J2 (and possibly J2J4). No toppling failure is possible for this geometry. Planar failure could occur along set J1 for fractures with dip angles smaller than the steep slope angle of the rock face (82°). Planar sliding along the fault considered as the basal plane (036/41) is also possible for this slope geometry.

For the case of the second slope geometry (central column in Figure 3.9c), planar sliding might be possible along the fractures parallel to foliation for a lower friction angle value than 26°. A reduced friction angle value for fracture sets parallel to foliation (So) is possible because several of the fractures are related to weathered shear zones, and in some cases, present fine grained infillings. Wedge failure is not kinematically possible for this geometry. Toppling is kinematically possible for this rock wall geometry along fracture set J4. Small toppled blocks bounded by J4 fracture planes have been observed during field recognition on this rock face.

For the case of the last rock wall geometry planar failure can occur along J2 and J4 fractures, J1 fractures in some cases can be affected by planar sliding too when they have a dip angle lower than the slope. The proposed basal surface can slide with the mentioned slope geometry. Wedge sliding can occur for the intersections J1J2 and J2J4 similar with what was the result for the first slope geometry. No toppling failure is possible for the proposed conditions.

Based on the kinematic analysis and field observations, toppling and planar sliding in direction SW are considered the potential failure mechanisms in compartment I. For compartment II and III, it is proposed that the whole mass slides along the basal surface. This asserts are supported by the results of the D-GPS monitoring network. Internal deformation is controlled by meso scale fractures of set J1, J2 and J4 through planar and wedge sliding towards the valley bottom causing the opening of those discontinuities as observed during field recognition.
3.4.3 Medji

The area of Medji, a ridge located to the SW of the village of St. Niklaus ranging from 1420 m to 2200 m asl., has shown intense recent activity and threatened people and infrastructure at the foothill and valley bottom. As presented in chapter 1 the lower part of the ridge (120,000 m$^3$) failed in November 2002 along a complex rupture surface (Ladner et al., 2004). The predominant lithology in the area is the Randa Augengneiss, slightly or no weathered. The gneiss is mostly weathered on the fracture surfaces by water circulation (infiltration). According with Joerg (2008) two main large discontinuities sets (fracture zones ST1: 020-040/90, and ST2: 020-040/40-65) corresponding to sets F2 and F4 on the present work (chapter 1) and two meso scale discontinuity sets (K1: 037/77; K2: 096/86) were found in the area that correspond with sets J3 (033/72) and J1 (100/81) in the present work (see chapter I). Foliation for the area has a mean orientation of 233/17 (240/18 according with Joerg, 2008). Small, meso and large discontinuities parallel to the foliation exist as well. These fracture sets correspond well with the sets presented on Ladner et al. (2004) as the main sets that control the failure in November 2002. Based on the analysis of geodetic data and remote sensing imagery before and after the event, a new interpretation of the mechanism that controlled this failure is presented here.

The failed rock mass was divided in four compartments covering an area of around 5100 m$^2$. The three lower compartments (II, III, IV in Figure 3.10a) are delimited by meso scale fractures (sets J3 and J1, Ladner et al., 2004) and a stepped basal failure surface following meso scale fractures from the set J1 and shear zones and fractures parallel to the foliation (See Figure 2.2, Chapter 2; Rovina, 2005). From geodetic monitoring before the failure event and inclinometer measurements inside a borehole drilled 4 months before the failure, it was determined deformation rates at the top of the unstable mass was larger than to the bottom, and several internal sliding surfaces were found at 6, 30 and 35 m deep. Vector displacement of the mass is 110/10-20 (Rovina, unpublished work). The main mechanism proposed for all three compartments in the present work is toppling with the basal surface located approximately 60 m deep. From the geodetic monitoring can be observed that direction of movement of those blocks is homogeneous and strikes to the east with increasing displacement rates up to the moment of failure (Figure 3.10a). The uppermost compartment (I) occupies approximately half of the total failed area. The lateral and back boundaries of this compartment are structurally controlled by meso scale fractures from sets J1 and J3, being its basal failure surface shallower respect to the other compartments. Planar sliding is being postulated in the present work as the main mechanism controlling failure of this compartment. The basal failure surface of compartment I had an orientation similar to the current topographic slope (120/20) and similar to the displacement vector calculated by Rovina (unpublished work) in that area. This plane
does not correspond to the orientations of fracture sets described for the western flank of the Matter Valley. The displacements observed on the geodetically controlled stations for this compartment strike to the southeast, and are higher than for the other compartments; similar to what occurred for other compartments, measurements show acceleration in the displacement rates close to the day of failure (Figure 3.10a).

Summarizing, the November 2002 rock slope failure in the area of Medji had a composed mechanism that divides the whole moving mass on 4 compartments delimited by meso scale fracture sets (J1, J2 and J3), the three lower compartments, representing half of the total area, failed by toppling along J1 structures and the remnant one (the uppermost compartment) by planar sliding along a fracture not belonging to any of the characteristic fracture sets for the area.

Several additional geodetic control points were installed after the 2002 event in the area that previously failed (Figure 3.10a). Some of them show ongoing displacements. However, field recognition showed that most of these stations are located on loose blocks. Additionally, after the 2002 failure at Medji, interferometry synthetic aperture measurements, both satellite based and ground based, were carried out (SLAM project, ASTRA project). Details of the methods were presented above. The results for the Medji area are shown in Figure 3.10b. Correlation between ground based and satellite based InSAR results cannot be executed because the GB InSAR data are not georeferenced, the data sets have different time frames and different line of sight direction. However, areas with evidence of ongoing displacements can be identified in both data sets. The mentioned results show deformation on an area wider than the area affected by the 2002 failure.

The results for both methods show higher deformation rates located above the area that failed in 2002. Results show different velocity ranges along the ridge and velocity compartments are delimited by lineaments corresponding to two large scale discontinuity sets (F2 and F4, see Chapter 1). Morphological features indicating signs of deformation (e.g.: scarps, former debris deposits) were identified and used to delimit the total extension of the moving area (shown as a blue line in Fig. 3.10b) covering approximately 250000 m². Five compartments are proposed as the most active areas in this moving mass (Figure 3.10b). The central area of the current unstable mass (which is located around the area of the 2002 failure) shows the most consistent displacement patterns in both data sets. Data from the geodetic network installed in the central part of the instability (Figure 3.10a) show a SE direction of displacement. However each compartment may have a different direction of displacement. The depth of failure and failure mode could not be assessed.
3.4.4 Sibulbodme

The rock slope instability at Sibulbodme is located on the upper part of Saenggini, on the southern flank of one of the secondary streams of the Matter Valley between 2320 m and 2410 m asl., on top of a small ridge dipping towards NE (Figure 3.11a, 3.11b). The instability covers an area of around 7000 m² and is located to the SE of a former rock slope failure of similar dimensions and apparently with a similar structural control (Figure 3.11a). The slope below the ridge is widely covered by an irregular layer of debris produced by recurrent small size rockfalls released from the toe of the mass, suggesting a high level of rockfall activity in the area. At some locations these deposits have been remobilized by small debris flows that have, in some cases, reached a headwater channel located deeper in the slope (as it can be inferred from photo interpretation of aerial photographs of the area). The Randa augengneiss is the main rock type for the unstable area with some narrow layers of paragneisses of the same composition as the Randa Augengneiss (Willenberg, 2004) located above, out of the unstable area. Strongly weathered mylonites are found on shear zones oriented parallel to foliation. Fractures corresponding to four meso-scale fracture sets are found within the unstable mass. J₁ (110/80), J₂ (063/74), J₃ (018/58), and J₄ (031/78). They are shown in the stereoplot presented on Figure 3.11b. Large structures from sets F₂ and F₃ have been observed controlling the limits of the unstable area.

The upper part of the instability is limited by a head scarp formed along a large vertical J₃ fracture (Chapter 1) striking NW–SE (Figure 3.11a). The same structure is the upper limit of the former rock slope failure mentioned before. Evidences of small size rockfall activity can also be observed above the head scarp from rock walls with a similar orientation. Both lateral flanks of the instability are structurally controlled by two large fractures (Figure 3.11b) from two different sets (a J₁ discontinuity to the NW and a F₃ discontinuity to the SE). The J₁ structure which is the NE limit of the current instability was the SW limit of the former rock slope failure in the area (Figures 3.11a and 3.11b). J₃ discontinuities control several secondary scarps in the mass (Figure 3.11b), which divide the rock mass in several compartments. Those compartments are furthermore segmented by J₁ discontinuities. Opening of discontinuities from both sets increases towards the toe of the instability indicating higher deformation rates for the downslope compartments. Small blocks located at the lateral boundaries of the unstable mass shows displacements towards the sides of the instability (perpendicular to set J₁). It was observed during field recognition that the base of the instability at the toe is located at the bottom of the ridge; around 25 m below the crown behind the headscarp. No direct observation of a large fracture at the base was possible. However the slope underneath the unstable blocks has a similar orientation than one of the large discontinuity set (F₂, average slope orientation 040/40) which is assumed to form the basal limit of the moving rock
mass. The toe of the instability coincides with a strongly weathered shear zone running parallel to the orientation of foliation (~230/20).

A D-GPS monitoring network was installed in the area comprising 18 stations surveyed two times with an interval of about one year. Locations for the stations was chosen trying to detect differences in direction of displacement of the compartments in the rock mass (Figure 3.11b). Results show that the rock mass moves homogeneously towards the NE, with higher displacements of about 20 mm/year in the lower section. The middle part of the instability moves with slower rates of displacement (average 15 mm/year). The upper part of the instability shows the slowest displacement rates (average 12 mm/year). The uppermost D-GPS stations were installed on the crown of the rock slope instability. No reliable values were measured (displacement rates smaller than 10 mm/year). Accuracy for the measurements is 10 mm.

Satellite based DInSAR results from the SLAM project were also available for this area. All targets are located above the rock slope instability. Apparent velocities range between -1.524 and 0.108 mm/year, having half of the targets velocity rates smaller than 0.1 mm/year. Simmons and Rosen (2007) report that for other studies with similar techniques (permanent scatterers), movement rates could detect variations with an accuracy of better than 1 mm/year. In the case of Sibulbodme, higher movement rates are to the NW of the unstable area and can be related with movement of loose blocks (targets are located on areas with shallow rockfall related debris deposits. Values at the crown of the instability are mostly smaller than 0.1 mm/year, shorter than the accuracy limit that can be obtained with the method. These targets are considered stable.

A schematic cross section of the rock slope instability is presented in Figure 3.11c. A combined mechanism of sliding-slumping controls the development of the instability. Fractures from set J4 and J3 correspond to the steep sets of the slumping blocks and form secondary scarps which segment the rock mass into several compartments. A planar basal surface controlled by a F2 discontinuity is proposed at the flat daylighting set for slumping. The main direction of displacement coincides with movement along the F2 basal surface. Lower compartments, closer to the toe, have larger displacements. Slumping mainly occurs in the compartments located in the lower part of the instability caused by a reduced friction angle along weathered shear zones with fine grained fillings. An example of the influence of shear zones is shown in figure 3.11d. The area in the figure is located to the toe of the rock slope instability.
3.5 Discussion and conclusions

A qualitative and semi-quantitative analysis of rock slopes failures has been carried out in the western flank of the Matter Valley, Switzerland, between St. Niklaus and Randa. A detailed inventory of the rock slope instabilities and associated deposits, using intensive field work, aerial photo-interpretation and DEM analysis was generated. Results show that the study area can be subdivided in two zones (northern and southern region) with significant differences in density and magnitude of rock slope failures and deposits. In addition three of the main large current instabilities were investigated in greater detail, including field mapping for structural characterization, and displacement monitoring with satellite based, and ground based InSAR, differential-GPS, and other geodetic methods. From these data sets comprehensive descriptions of large scale failure mechanisms have been derived. Details of these investigations are discussed below.

3.5.1 Landslide phenomena in the Matter Valley

The western flank of the Matter valley between St. Niklaus and Randa has a remarkable topography composed of a valley bottom with substantial soil infill at an elevation of 1100 to 1200 m asl, followed by a steep glacial slope (typically 70º) up to an elevation of about 2000-2200 m asl. This glacial slope is dissected by side valleys with perennial streams and numerous dry gullies. These gullies follow large scale faults (see Chapter 1). Above the flat shoulder at 2000-2200 m asl elevation follows a complex topography up to an elevation of 3500-3800 m asl with a mean slope angle of typically 40-50º.

Many small and large size rock slope failures have occurred in the past causing debris accumulation at the toe of steep rock slopes along the valley bottom and side valleys, and on terraces of stepped slopes. Characterization of the different type of deposits was done based on their morphologic characteristics with the help of high resolution DEMs. The majority of the deposits are located at the toe of steep slopes of the main valley. Only few deposits are found outside of the main valley in upper parts of the study area. Most of the debris produced by rock slope failures in the side valleys has been transported to the main valley by multiple episodes of debris flows that with time have generated debris flow cones at the outlets of the lateral streams to the main valley.

In addition, shallow debris deposits cover flatter areas on the shoulders of stepped slopes on the flanks of the lateral valleys. These deposits have probably formed by a succession of small rockfall events with short runouts that did not reach the lateral streams. An example of this type of deposit is found at Sibulbodme where a large area is covered by debris deposits. These deposits were
probably formed by rockfalls with a similar structural control and failure mechanism as the ones that control the current instability in the area.

Intermediate and large scale rock slope failures (release areas) have been systematically mapped and classified based on their age and size. Small scale failures of individual blocks were only used to further classify stable rock slopes based on the density of these small events identified on 3D models created by ground based photogrammetry (a detailed analysis of the small rockfall failures is given in the next chapter). The larger rock slope failures have been grouped into lateglacial and postglacial according with their age. It is assumed that the valley glaciers during the older Dryas (lateglacial) covered the Matter Valley bottom as far as to St. Niklaus based on the observations presented for other glacial valleys in the region with similar characteristics (Ivy-Ochs et al., 2008). It is assumed that deposits of rock slope failures right after LGM were washed out by the new advance of the glacier during the lateglacial stadials. Large rock slope release areas without corresponding deposit are therefore classified as lateglacial, and release areas for which the corresponding deposits are identified are classified as postglacial. Besides this relationship also the weathering characteristics of the release surfaces have been considered to get a relative age of the rock failures. Lateglacial release surfaces have been identified on the steep flanks of the main valley or in SW facing side valley slopes. With the exception of Randa, postglacial failures are preferentially also located on the steep walls parallel to the main valley but are in average smaller than the late glacial failures.

Based on the spatial distribution and size of colluvial deposits the study area can be divided in two regions separated by the Blattbach River (Coordinate 110900). At this location the axis of the Matter Valley rotates about 15°. Both domains also indicate a different geomorphological evolution which is expressed by differences in the frequency, depth and width of the incipient gullies eroded into the steep sidewalls of the western Matter valley.

In the southern domain, many lateglacial release surfaces have been found and they are mostly located on the front walls of the main valley with no remnant deposit. In this domain, magnitudes of the deposits are comparatively small for all types of deposits. Rock fall debris cones occupy areas that range from few thousands to few tens of thousands of square meters. They can be delimited individually because no coalescence exists between deposits. In several cases debris cones can be directly related to the location of mapped (postglacial) release areas. At the southern boundary of the study area (close to Randa), rectilinear talus slopes control the type of deposition implying only an incipient segmentation of the rock slopes. Towards the north, slopes are increasingly more bisected by incipient linear gullies and a larger number of debris cone deposits close to the Blattbach River. Linear erosion along incipient gullies (expressed by the depth and width of incision) increases towards the north, and causes an increase in the magnitude of the debris cones and the
presence of coalescent deposits. However, the length of the gullies stays in the same order as in the south of the study area. Deposits related to large rock slope failures have not been found in the southern region due to the wash out of the deposits during more recent glacial stadials.

In the northern domain, magnitude and frequency of large rock slope failures, expressed by the magnitude of their debris cones, is larger than in the south, creating deposits with magnitudes in the order of several tens of thousands of square meters. Most (postglacial) slope failures with genetically correlated deposits are located in the steep walls of the main valley slopes and a few in the southwestern flanks of the lateral valleys. Rock failures with no deposits in this domain are mainly located close to its southern boundary (Blattbach River).

Also the current intensity of recent small rockfalls is comparatively higher in the north than in the south documented by several un-vegetated slope patches that can be mapped both on debris cones and alluvial fans using aerial photographs. 3D models created by ground based photogrammetry show higher density of former source areas than in the south. Larger debris flow cones are as well observed in this domain. They have an important component of debris contributed by several other types of mass wasting processes (e.g. snow avalanches, debris flows) that makes difficult to estimate the frequency of only single small rockfall events. In the northern domain rock walls are more densely bisected resulting in coalescent cones, except in the area of Sparru, where single cones can still be observed. Linear incision of the incipient gullies is deeper and wider, which greatly modify the slope shape (e.g. the area of Walkerschmatt bisected by two gullies with large debris cones at their bottom). Deposits product of large rock slope failures are interlaced with debris cones and have been later covered by debris product of more recent small rockfall events making difficult to isolate single deposits.

The northernmost slope in the area, Sparru, presents special characteristics that do not fully correspond to observations of the other slopes in the area and especially to the observations of other slopes in the northern domain. Neither coalescent debris cones exist nor deep linear incision along gullies is observed. Large rectilinear talus slopes appear between debris cones which magnitudes larger than the ones observed at the southern boundary of the study area. These differences might be related to changes in lithology (schists and quarsites in contrast to the predominance of the augengneiss in the other regions), and to higher erosion of the Matter-Vispa River in this part of the study area. At this location, the valley shows a transition from glacial (U-shaped) to fluvial (V-shaped) caused by an increase in fluvial incision (Leith, personal communication).
### 3.5.2 Current rock slope instabilities in the Matter Valley

Three of the larger active instabilities were studied in detail: Walkerschmatt, Medji and Sibulbodme. These instabilities are located in different morphological settings and broadly share similar structural and lithological characteristics with site particularities. Every site presents a different behavior and is controlled by different mechanisms. Regional fracture sets, both meso and large scale, presented in chapter 1 of this work, play an important role on the behavior of these instabilities.

For the case of the rock slope instability at Walkerschmatt, based on the preferential mechanism two domains can be established:

At the southern cliff of the instability the main mechanism for active block release is toppling. Toppling occurs on single blocks laterally controlled by fractures corresponding to sets J1 and J4 and with a basal surface of fractures and shear zones parallel to foliation. Blocks topple towards the S-SW, perpendicular to the WNW-ESE striking steep cliff. In some cases depending on the geometry between the rock slope, foliation, and block shape some blocks can slide along fractures parallel to the foliation. At the toe of this cliff, a shallow debris deposit is located evidencing the recurrence of small rockfall failures.

At the northern domain planar sliding along fractures J1, J2 and J4, and wedge sliding to the N-NE along the intersections J1J2, and J2J4 are kinematically possible. The domain is limited by two steep rock walls to the E (NE-SW cliff) and N (ENE-WSW cliff). Deformation causes fracture opening along set J1 and J2. The northern domain has been divided in two compartments, with the eastern compartment (compartment II) considered more active based on fracture openings, ambient vibration test results, and D-GPS measurements.

For the case of the rock slope instability at Sibulbodme a sliding-slumping mechanism is proposed. The whole rock mass is sliding along a F2 plane parallel to the down-slope topography with the whole rock mass separated in several compartments controlled by J3-J4 fractures. Displacements of the compartments measured by D-GPS are larger towards the toe of the rock mass causing a larger displacement along active fractures. Slumping is mainly observed at the toe of the instability and it is caused by strength degradation caused by the presence of heavily weathered shear zones parallel to foliation.

The failure of November 2002 at Medji occurred by toppling of the lower part, divided in three compartments limited by J1 meso-scale fractures, and sliding of the upper part along a plane parallel to the current topography (120/20, parallel to the calculated displacement vector) not corresponding to any of the fracture sets found in the area. The upper compartment showed higher displacement
rates before failure and a different direction of displacement than the rest of the rock mass. Failure of the lower compartments may have been a consequence of the sliding of the upper compartment that pushed the three lower compartments causing a change in their local stress conditions. A larger instability in the area Medji has been detected by SB- and GB-InSAR with displacements observed on the slope up to elevations of 1900-2200 m asl. This deformation can be correlated with observed morphological features (e.g.: open cracks, scarps). The rock mass is divided in five compartments delimited by morphological features (lineaments) identified on aerial photographs that follow similar orientations as the regional large fracture sets (see Chapter 2).

Intermediate and large current rock slope instabilities are widespread on the study area and are not specifically concentrated in the southern or northern domain. The instabilities occur at different heights (from ~1800 to ~2400 m asl), mainly on steep slopes facing to the S or SE. A special case is the area of Medji, the largest active instability detected in the area, where both flanks of an east-trend ridge are part of the unstable area.

The slope geometry has a relationship with the type of mechanism controlling failure. E-NE facing slopes are prone to fail by sliding (northern compartment Walkerschmatt, Sibulbodme), whereas SE-S facing slopes are more susceptible to fail by toppling (southern compartment Walkerschmatt, Medji rockslide in 2002). Sliding occurs along pre-existent, moderately steep fractures (set F2 for the cases above mentioned) on either moderately steep or steep slopes. Toppling is controlled by steep fractures (mainly J4, with set J1 as lateral boundaries) with foliation parallel fractures acting as the basal planes on steep rock faces.

For the bulk rock mass at Walkerschmatt (northern compartments) planar sliding is most probably the dominant mechanism. At Sibulbodme, sliding and slumping have been identified as major mechanisms. The 2002 failure of Medji and the current displacements at Randa are controlled by two mechanisms: toppling and sliding; each mechanism controlling a separate section of the moving rock masses. In most cases, internal deformation of the rock masses are controlled by steep meso-scale fracture sets, and large discontinuities (faults) control the boundaries of the instabilities (e.g. Sibulbodme, Medji, Randa).

D-GPS measurements in the selected areas have helped to understand the deformation style of the unstable rock masses, however a longer period of monitoring is recommended to have a clearer comprehension of the kinematics. InSAR techniques were useful to monitor slow displacements in the instabilities but targets inside of the most active areas of the instabilities were scarce, the latter shows the importance of field control and a previous knowledge of the areas before target selection.
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Figure 3.1. Map of susceptibility for rock slope instabilities for the area of the Matter Valley. The largest debris cone to the south corresponds to the 1991 Randa rockslide events.
Figure 3.2. a) Aerial photograph showing characteristic examples of the three main classes of deposits at the study area of Randa; b) Mapped deposits overlaying a 3D shaded relief map. It is clearly observable that several additional features of the deposits can be easily mapped on the latter. A, A', B, B', C and C' show the location of the topographical profiles for the three types of deposits shown on c) for the debris cone example d) for talus slope and e) for alluvial fans. The profiles shown both the heights (solid black lines) and the slope angle values (solid blue lines). Values extracted from a 2.5 m DEM.
Figure 3.3. Histograms showing a summary of the main geometry characteristics (area, slope and length) for the main three types of deposits found on the study area: a) debris cones, b) alluvial fans and c) talus slopes. Values are extracted from a 2.5 m DEM based on the landslides inventory map.
Figure 3.4. Example of a moraine deposit on an hanging valley in the western flank of the Matter Valley to the south of Walkerschmatt a) Characteristic texture as observed on a 3D relief map, b) overview of the hanging valley from the east, in the foreground to the right, part of the moraine deposit. In the center, levees product of recent debris flows, c) Transversal NE-SW profile across the deposit showing its geometrical characteristics. The sudden raise on the slope angle values to the south is due to the presence of the levees close to the main stream.
Figure 3.5. Composite deposit in the area of Medji: a) 3D relief map showing the contour of the deposit; b) Photograph taken from the NE, showing a partial view of the deposit; c) Profile A – A’, crossing one of the two long lobes to the north; d) Profile B – B’ along the steepest part of the deposit to the south.
Figure 3.6. Debris on slope example located on the area of Seemate. a) Aerial view showing two different shallow deposits at two different levels on the stepped slope (see contour lines); b) Photograph of the same area taken from the south east. In the upper part of the photograph the toe of the upmost deposit is visible, several blocks in the proximity of the main deposit product of small size rockfalls.

Figure 3.7. Small size recent failures observed on a 3D visualization from a DEM generated by ground based photogrammetry. a) location area, purple line encloses area surveyed by ground based photogrammetry; b) 3D view showing overhanging blocks that disguise the morphological expression of release areas on aerial based elevation models.
Figure 3.8. Planimetric view of the D-GPS network implemented on the western flank of the Matter Valley.
Figure 3.9a. Structural setting and D-GPS and geodetic monitoring networks of the Walkerschmatt rock slope instability. On the upper right corner, schematic representation of the failure mechanism.
Figure 3.9b. Ambient vibration results and compartments of the Medji rock slope instability.
Figure 3.9c. Kinematic analysis for the area of Walkerschmatt. Stereoplots to the left are built for a rock face orientated 115/82, in the center show results for a wall with an orientation of 190/63. To the right for a rock facing 080/60. 34 orientation values were used for the analysis clustered in sets.
Figure 3.10a. Displacement rates of 3D geodetically controlled points in the area of the 2002 Medji rockslide before and after the event and structural compartments of the rock mass involved in the failure. On the upper right corner, displacement rates of selected stations.
Figure 3.10b. Overview of the current rock slope instability at Medji. Colored polygons represent the state of activity of different compartments based on ground based and Satellite based InSAR data (LISALAB, 2007; SLAM project). The blue polygon shows the limit of the instability. Colored dots shows yearly velocity rate of displacements from satellite based DInSAR.
Figure 3.11a. Overview of the Sibulbodme rock slope stability showing the main morphological features. Picture taken from NE.

Figure 3.11b. Displacement direction and rates from D-GPS monitoring at Sibulbodme. On the upper left corner, stereoplot showing kinematic analysis results for planar sliding. On the lower left corner, an example of an open J3 fracture product of sliding. The cross section A-B is shown in Figure 11c.
Figure 3.11c. Schematic cross section of the rock slope instability at Sibulbodme.

Figure 11d. Sketch showing a clear case of slumping at the toe of the Sibulbodme rock slope instability.
Chapter 4


4.1 Introduction

Rockfalls are a widespread hazard in steep crystalline rock slopes. In such rock types, fractures play a key role controlling the type and volume of failed rock masses (Piteau, 1972; Grenon & Hadjigeorgiou, 2007). For coherent blocks of simple geometry three main mechanisms control failure processes: plane sliding, wedge sliding and toppling (Hoek & Bray, 1981). In addition slumping can been considered as a fourth failure mechanism, which is not included in standard slope failure analyses. Kinematic analysis is often used to assess the possibility that a rock slope failure with simple geometry can occur, or that for some geometry of slope and discontinuity orientation movement is possible (Hudson & Harrison, 1997). Traditionally, the kinematic analysis is done using deterministic methods and representative values of orientation and strength parameters. However, all fracture properties have a significant level of scattering (Feng & Latjai, 1998). Moreover, natural rock slopes are not geometrically regular; their aspect widely varies even for small areas. A schematic representation of these conditions is shown on Figure 4.1.

A common approach to deal with the uncertainty caused by the variability of discontinuity set orientations is applying probabilistic methods for slope stability analysis. Monte Carlo simulations are often used and consist of numerical methods that solve mathematical problems through random sampling of variables (Feng & Latjai, 1998). Several authors have developed computer programs and codes to model the afore mentioned variation for plane sliding (Singh et al., 1985), wedge sliding (Quek & Leung, 1995; Latjai & Carter, 1989; Piteau et al., 1985) and toppling (Tatone & Grasselli, 2010, Scavia et al., 1990). These methods have become a standard tool for hazard analysis. However, most applications have been used for specific site investigations with defined slope geometries not taking in account spatial variability of slope angle and aspect along the rock surface as it is often the case for natural rock slopes. In fact, only few attempts exist that apply kinematic stability analysis for a regional investigation, and they have used a deterministic approach...
(Jabodyoff, 2002; Günther, 2003; Kim et al., 2004) or a discrete series of data using fuzzy logic (Aksoy & Ercanoglu, 2007).

An important constraint for the development of regional rock slope failure analysis methods in the past was the lack of accurate terrain models. In recent years, the development of new techniques for digital elevation model (DEM) generation, like airborne and ground-based LiDAR and ground-based photogrammetry, has resulted in an increasing accuracy, as well as in an improvement of spatial resolution. In addition ground-based photogrammetry can be used as a tool to validate the results of kinematic and slope analysis by visual checking of release areas.

Geographic information systems (GIS) are regularly used to analyze spatial information and to integrate information from different sources in consolidated and easy-to-query databases. GIS have been widely used for landslide analysis in several ways e.g. for landslide susceptibility mapping (Günther et al., 2004), rock fall run-out assessments (Frattini et al., 2008), and the identification of rock fall source areas (Loye et al., 2009). Probabilistic methods combined with GIS have been developed mainly for landslide hazard susceptibility of soil failures. There have been only few areal studies based on kinematic analysis principles (i.e.: Jaboyedoff et al., 2004; Kim et al., 2004, Derron et al., 2005).

For the present work, stochastic methods considering fracture properties as random variables were implemented in a GIS environment in order to assess the stability of natural irregular rock slopes using principles of kinematic stability analysis. The methodology is developed for small rock slope failures (volumes in the order of cubic meters and tens of cubic meters).

The study area corresponds to the western flank of the Matter Valley, Switzerland between the localities of Randa and Sankt Niklaus. The area consists mainly of Randa Augengneiss, a metamorphosed Permian porphyritic alkaline to subalkaline granitic intrusion constituting the core of the Penninic Siviez-Mischabel nappe (Willenberg, 2004) dipping gently towards the southwest. The current morphology reflects Pleistocene glacial erosion, which led to the formation of a U-shaped valley overprinted by postglacial gravity-driven processes. The western face of the valley is characterized by steep rock cliffs carved in the Augengneiss that, because of their high relief (in some cases beyond 1000 meters), have been recurrently affected by rock fall events (see chapter 3).

### 4.2 Structural characterization of fracture sets

For structural characterization of the study area a comprehensive dataset of fracture geometries was created by field work and DEM analysis. The area was divided in ten domains, each one representing E-W ridges that form the western flank of the Matter Valley (See figure 2.1 on chapter
2). Only meso-scale discontinuities were used, considering that most of the rock slope failures have volumes in the order of tens of cubic meters and larger. Meso-scale discontinuities are defined as rock breaks where only minor tectonic displacement has been found and which possess persistence values in the order of meters to tens of meters. An example of such discontinuities is given in Figure 4.1. Approximately 2300 measurements of meso-scale discontinuities were collected in the entire area during field work.

A considerably larger amount of data was indirectly extracted from an airborne-based LIDAR high resolution DEM (HR DEM) provided by SWISSTOPO with a spatial resolution of 2.5m. This was possible for the study area because the fracture pattern has a strong influence on the morphology of the steep rock slopes and vegetation is often scarce. To measure the orientation of slope faces controlled by fractures, a 3D shaded relief map was produced. A 3D shaded relief map is a color-coded image based on HSV (Hue-Saturation-Value) color composition showing changes in color according with the changes in slope orientation. Saturation is represented by the dip angle (slope). Hue corresponds to the dip direction angle (aspect). Value for this case is assumed to be constant and equal to 90. Domains for fracture orientation analysis were carefully selected taking in consideration that not all pixels represent fracture planes. Selection of pixels was done using 3D visualizations, and an orthophoto mosaic created with aerial photographs acquired in 2005 (SWISSTOPO) was used as the top-most layer. The analysis was also supported by photographs of the slopes taken from different angles during field work. Finally, an area correction was performed in a similar way as Terzaghi’s correction to avoid the bias produced by the ratio between number of pixels and slope angle. A total of more than 120000 fracture orientations were obtained from this analysis (chapter 2).

Clustering of the fracture orientations was studied with Fisher distributions using the code DIPS (Rocscience). Field measurements and DEM-based data were processed separately. Five main systematic fracture sets have been found in the study area (Table 4.1), which varies in orientation gradually from the north to the south. For these fracture sets, an additional comparison between field work and DEM analysis was done using both components of fracture orientations (dip and dip direction) and to integrate both types of results. All values of fracture orientation that correspond to a fracture set cluster were used to assess their probability density function (PDF). It was found that all datasets fit well with a normal-type distribution. Using the parameters for normal distributions (mean and standard deviation), a multiple-sample method for statistical inference called ANOVA (analysis of variance) was used to compare the statistics of field measurement and DEM-based values. In this method, differences among several populations are regarded as errors. In previous geological studies, ANOVA was used for the comparison of field data from different sites. In our
case, field data was compared to DEM-based data domain by domain, showing a good fit with index
values ranging from 0.006 to 1.57 (good fit). A thoroughly description of results and methods are
included in the chapter one of this work.

4.3 Kinematic analysis using three-dimensional vector algebra.

Principles of linear algebra were used to represent the conditions needed for failure to occur. A
simple introduction to the principles and main definitions used in this work follows:

**Plunge (-90° ≤ β ≤ 90°):** Acute angle measured in a vertical plane between a given line and the
horizontal plane. Downward sense means positive plunge.

**Trend (0° ≤ α ≤ 360°):** The geographical azimuth, measured in clockwise rotation from north (0°)
of a vertical plane containing the given line; trend is the azimuth that corresponds to the direction
of plunge of a line.

The trend (α) and the downward plunge (β) of the line of maximum dip (plunge) of a plane are
referred to as the Dip direction and the Dip angle of such a plane.

For a line located at right angles to a given plane (normal to a plane):

\[ \alpha_n = \alpha_d \pm 180° \text{ and } 0° \leq \alpha_n \leq 360° \]  
\[ \beta_n = 90° - \beta_d \text{ and } 0° \leq \beta_n \leq 90° \]

<table>
<thead>
<tr>
<th>Set</th>
<th>Dip</th>
<th>Dip direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>63 - 83</td>
<td>097 - 126</td>
</tr>
<tr>
<td>2</td>
<td>59 - 78</td>
<td>66 - 80</td>
</tr>
<tr>
<td>3</td>
<td>45 - 79</td>
<td>015 - 037</td>
</tr>
<tr>
<td>4</td>
<td>76 - 83</td>
<td>165 - 192</td>
</tr>
<tr>
<td>5</td>
<td>02 - 42</td>
<td>105 - 336</td>
</tr>
</tbody>
</table>

*Table 4.1. Range of orientations of the five main fracture sets found in the study area. Set 5 corresponds
to a fracture set parallel to foliation.*

**Strike (α):** Trend direction of a horizontal line in a given plane. The strike direction (0° to 180°)
corresponds to the strike of the plane. For a given plane the strike will lay 90° from \( \alpha_n \) and 90° from
\( \alpha_n \).
We used a right-handed Cartesian system with north pointing in the positive y-axis. For a Plane U (Figure 4.2) and a unit vector \( \mathbf{u} \) following its dip direction:

\[
|\mathbf{u}| = \sqrt{u_x^2 + u_y^2 + u_z^2}; \quad |\mathbf{u}|: \text{magnitude of U} \quad (4.3)
\]

\[
\alpha = \tan^{-1}\left(\frac{u_x}{u_y}\right) + Q_1; \quad \alpha: \text{dip direction angle of U} \quad (4.4)
\]

\( Q_1: \text{correction factor for } \alpha \)

\[
\beta = \tan^{-1}\left(\frac{u_z}{\sqrt{u_x^2 + u_y^2}}\right); \quad \beta: \text{dip angle of U} \quad (4.5)
\]

\[
\alpha_s = \tan^{-1}\left(\frac{u_y}{u_x}\right) + Q_2; \quad \alpha_s: \text{strike of U} \quad (4.6)
\]

\( Q_2: \text{correction factor for } \alpha_s \)

Correction factors \( Q_1 \) and \( Q_2 \) on equations 4.4 and 4.6 are required to obtain results over the range of 0° to 360°. Values of the equations without correction are in the range of ± 90°. Correction factors are in the range of 0° to 360°, and are related to the quadrant in the Cartesian system that the unit vector is placed.

The direction cosines of \( \mathbf{u} \) are:

\[
u_x = |\mathbf{u}| \sin \alpha \cos \beta \quad (4.7)
\]

\[
u_y = |\mathbf{u}| \cos \alpha \cos \beta \quad (4.8)
\]

\[
u_z = |\mathbf{u}| \sin \beta \quad (4.9)
\]

\( x, y, \text{and } z \) are sub-indices corresponding to the position of the direction cosines aligned to one of the axes of the Cartesian system.

The normal vector to \( \mathbf{u} \) is:

\[ \mathbf{n} = (\sin \alpha \sin \beta, \cos \alpha \sin \beta, \cos \beta) \quad (4.10) \]

For the case of the intersection of two planes (\( \mathbf{l}_{12} \)), the cross product of their normal vectors corresponds to their line of intersection (Groshong, 2006).

\[
\mathbf{l}_{12} = \mathbf{n}_1 \times \mathbf{n}_2 = \begin{pmatrix}
\cos \alpha_1 \sin \beta_1 \cos \beta_2 - \cos \alpha_2 \cos \beta_1 \sin \beta_2 \\
\sin \alpha_1 \cos \beta_1 \sin \beta_2 - \sin \alpha_2 \sin \beta_1 \cos \beta_2 \\
\sin \alpha_1 \cos \alpha_1 \sin \beta_1 \sin \beta_2 - \cos \alpha_2 \sin \beta_1 \sin \beta_2
\end{pmatrix} = \begin{pmatrix}
l_1 \\
l_2 \\
l_3
\end{pmatrix} \quad (4.11)
\]
Being $l_i$, $l_j$, and $l_k$ the coordinates of a vector parallel to the line of intersection $l_i$, in terms of its direction cosines.

The orientation of the intersection line is:

$$\alpha_j = \tan^{-1}\left(\frac{l_j}{l_i}\right); \quad \alpha_j : \text{direction of intersection line} \quad (4.12)$$

$$\beta_j = \sin^{-1}\left(\frac{|l_k|}{\sqrt{l_i^2 + l_j^2 + l_k^2}}\right); \quad \beta_j : \text{dip of intersection line} \quad (4.13)$$

Three mechanisms were analyzed: plane and wedge sliding, and toppling. The conditions implemented are based on the conditions presented in Hudson & Harrison (1997). The principles and their mathematical representation are presented below for each case. Other types of mechanisms (i.e.: slumping) were not analyzed.

**4.3.1 Single plane sliding**

According to Hudson & Harrison (1997), there are 4 conditions to be accomplished for slope plane sliding (Figure 4.3):

i. The dip of slope must exceed the dip of the potential slip plane.

ii. The potential slip plane must daylight on the slope plane.

iii. The dip of the potential slip plane must be such that the strength of the plane is reached.

iv. The dip direction of the sliding plane should lie within approximately $\pm 20^\circ$ of the dip direction of the slope.

These conditions are fulfilled by the following expressions (Kim et al., 2004):

$$\theta^' > \beta > \phi$$

$$\theta^' = \tan^{-1}\left(\tan \theta \cos \gamma\right)$$

Where:

$\theta^'$: apparent dip of slope respect to the discontinuity

$\theta$: slope dip

$\phi$: friction angle of discontinuity

$\gamma$: angle between dip direction of slope and dip direction of discontinuity
Conditions i, ii, and iv are accomplished when the apparent dip of the discontinuity at its intersection with the dip of the slope is greater than the dip of the slope (first and second member of eq. 4.14). Condition iii is fulfilled when the dip angle of the discontinuity is larger than the friction angle of the rock slope (second and third member of the eq. 4.14).

### 4.3.2 Wedge sliding

Three conditions have to be accomplished for wedge sliding (Hudson & Henderson, 1997). Schematically shown in Figure 4.4:

i. The dip of the slope must exceed the dip of the line of intersection of the two discontinuity planes associated with the potentially unstable wedge.

ii. The line of intersection of the two discontinuity planes associated with the potentially unstable wedge must daylight on the slope plane.

iii. The dip of the line of intersection of the two discontinuity planes associated with the potentially unstable wedge must be such that the strengths of the two planes are reached.

Mathematically these conditions are expressed by (Kim et al., 2004):

\[ \theta > \alpha > \phi \]  \hspace{1cm} (4.16)

\[ 270^\circ < \gamma < 360^\circ \lor \gamma < 90^\circ \]  \hspace{1cm} (4.17)

Where:

\[ \gamma : \text{angle between dip direction of the slope and dip direction of the line product of the plane intersection.} \]

Condition i is evaluated by first and second member of eq. 4.16. Condition ii is evaluated by eq. 4.17. Condition iii is expressed by second and third member of eq. 4.16.

### 4.3.3 Single block toppling

Two conditions to be accomplished for block toppling failure (Figure 4.5). Conditions presented here are based on Hudson & Henderson (1997) applied for the case of a block delimited by 3 fractures set (2 on the sides and 1 at the base):

i. There are two sets of discontinuity planes whose intersection dips into the slope.

ii. There is a set of discontinuities to form the base of the toppling blocks.
iii. Dip of the basal plane must be less than friction angle of the rock slope (to avoid occurrence of a sliding component for the case when the basal plane dips out of the slope).

These conditions are mathematically expressed by (adapted from Kim et al., 2004):

\[
150^\circ < |\beta| < 210^\circ \quad (4.18)
\]

\[
\theta > (90 - \beta_b + \phi) \quad (4.19)
\]

\[
\beta_b < \phi \quad (4.20)
\]

Where:

\( \beta_b \): Dip angle of basal plane.

Condition i is accomplished when eq. 4.18 is fulfilled; condition ii is fulfilled by eq. 4.19; and condition iii by eq. 4.20.

### 4.4 GIS implementation of kinematic probabilistic analysis

A Monte Carlo simulation has been implemented using Model Builder into ArcGIS v. 9.3 and later scripted on Python. Monte Carlo methods simulate a phenomenon which is based on the distribution of one or several variables in a mathematical model. A script for every type of mechanism was written using the conditions presented above. An example of the code's structure is presented in Figure 4.6 for the case of plane failure; the other two cases follow a similar logic.

For the analysis, dip and dip direction of the discontinuity sets and friction angle values were designated as to have a random behavior. It has been demonstrated that these parameters follow probabilistic distributions (Tran, 2007), commonly Weibull and/or normal distributions. For the present study, an intensive analysis of data corresponding to dip and dip direction of discontinuity sets in the area of the Matter Valley, Switzerland was carried out based on field work and DEM interpretation, as described above. This analysis confirms that data for the area fit well with normal distributions (see chapter 2). ArcGIS software has built-in tools that allow the generation of raster datasets following several types of probability distributions including normal distribution.

Friction angle was assumed to follow a normal distribution and the calculation of the distribution parameters, mean and standard deviation was based on literature data. According with Priest (1993) friction angle varies between 23° and 29° for similar materials as the Randa Augengneiss. A mean value of 26° and a standard deviation value of 1 were used to recreate those boundaries. The distribution of friction angles values was considered being the same for the entire
study area (considering that it is composed of homogeneous granitic gneisses, i.e. the Randa Augengneiss) and all fracture sets.

In this study slope geometry parameters have been considered constant for each DTM pixel of 2.5 m. However, this slope geometry contains a certain degree of inaccuracy due to the fact that the generation of a continuous grid of values is done by interpolation of discrete data. The problem has been discussed by several authors (e.g.: Raaflaub, 2006). The parameters considered for the present analysis are slope angle and slope aspect which are calculated using standard algorithms. Slope is defined as a pixel value that represents the rate of change of the surface in the horizontal and vertical directions from a center cell in a 3x3 moving window. Aspect is defined as the direction of the central pixel in a 3x3 moving window with respect to the north. The HR DEM from aerial LIDAR was used for the analysis. This DEM covers the whole area below 2000 m a.s.l. which corresponds well with the steeper part of the slopes where most of the smaller rock slope failures have occurred in the past.

The GIS implementation uses a cell-by-cell basis, evaluating the conditions for occurrence of failure pixel-by-pixel through a repetitive process, each new run producing newly sampled raster datasets for the fracture sets orientation and friction angle. Every time a pixel is considered as prone to fail, a value of one is assigned to the pixel in the resultant raster dataset, producing a raster image with values of zero or one each time. When the maximum number of runs is reached, a sum of all resulting datasets is executed using map algebra. Probability of failure is calculated by the number of hits a pixel has received during the process.

4.5 Ground based photogrammetry

Ground-based (GB) digital photogrammetry has been used in several fields of application such as architecture, civil engineering, conservation of monuments, archaeology, industrial photogrammetry, mobile mapping, or computer controlled navigation (Kraus, 2007). It has started to become a popular tool for structural and rock mass characterization of natural and human excavated rock slopes (Kemeny & Post, 2003; Haneberg, 2008) in mining and engineering geology projects. It presents some advantages with respect to other traditional field methods (e.g. scanline and window mapping) because of a comparatively larger area of surveying, larger data collections, shorter collection time and higher levels of safety for the surveyor. In recent years, GB photogrammetry has started to be used for other geological purposes, being rock slope instability analysis one of the most promissory fields of application (Sturzenegger & Stead, 2009a).
Photogrammetry is based on the use of overlapping sets of photographs for the generation of 3D surfaces and rectified photographs (orthophotos). The process is based in a maximum of 9 variables which resolve the internal and external orientation of cameras. For the case of digital GB photogrammetry applied for rock slope problems, commercial software based on robust algorithms and off-the-shelf cameras (e.g. ShapeMetrIX3D, Sirovision, 3DM Analyst) have shown to produce reliable results with sufficient accuracy because of the great control that the surveyor exerts during the data collection. Haneberg (2008), shows that using a digital camera Nikon D200 with a spatial resolution of 10.2 Mpixel and 20 mm focal length it is possible to produce a model with a point spacing of about 25 mm from a distance to the rock wall of approximately 20 m, what is comparable with results obtained using active sensors (i.e. LiDAR systems). Sturzenegger & Stead (2009b) point out that both methods (LiDAR and GB photogrammetry) present more integral results for discontinuity analysis on rock slopes than traditional manual surveys because they cover wider surfaces, giving a more representative view of the whole outcrops (lower part of slopes may present different discontinuity characteristics than upper parts).

The most common information extracted from terrestrial remote sensing methods for rock slope characterization is related to discontinuity characterization. Several works have shown the capabilities of those methods to obtain information of different discontinuity properties (e.g. orientation, persistence, roughness). Most of the work has been directed towards the extraction of orientations (Lato et al, 2009; Haneberg, 2008). Discontinuity orientation is extracted using vector algebra (direction cosines of the normal vector to the desired plane). Usually the selection of the planes is done by manual or semi-automated techniques. An advantage of digital photogrammetry over other remote sensing methods is the generation of 3D views with draped orthophotos of the surveyed rock slopes. These models allow to the user to make a more accurate selection of the fracture planes. A disadvantage is that, as a passive method, it is strongly influenced by external conditions (e.g. light conditions).

Models can be referenced by positioning the acquisition points (camera position) or by positioning target points on the imaged rock slope. Models can be locally or globally referenced. For a global referencing, global positioning systems (GPS) are often used. In the case of referencing using target points on the slope, a range finder (e.g. Leica Laser Locator) or a total station are commonly used. Targets can be surveyed from any position. For a global referencing using target points an accurate coordinate of the surveying location is needed. To improve the accuracy in the global referencing of the models, differential GPS (D-GPS) techniques are commonly applied.

For the present work ground-based photogrammetry was used to test the validity of the GIS-based kinematic failure analysis. For this purpose, areas of the rock slopes where evidence of past
failure events existed were selected. The photogrammetric models were georeferenced using natural targets such as intersection of discontinuities, which were surveyed by a reflectorless total station (Leica system 1200) whose position was obtained using differential GPS systems (GPS Leica system 1200). 3D visualizations from ground-based photogrammetric elevation models were used to digitize release areas of former events and extract the orientation of their slip surfaces to compare with orientations of the fracture sets found in the structural analysis (Chapter 1). The results from ground based photogrammetry and the stochastic models were compared using ArcGIS. An example of the results from ground based photogrammetry is presented in Figure 4.7. The stereoplot shows that, for the area of Guggini, orientations of the failure surfaces follow planes with orientations similar to sets 1 and 2 which confirm a structural control in small size rock slope failures of the main fracture sets found in the area.

4.6 Results and discussion

Several datasets have been generated for each structural domain of the study area, and the different fracture set combinations and mechanisms e.g.: wedge sliding failure was assessed with fracture sets 1 and 2, plane sliding with set 1, toppling with fracture set 2 and 3 as block walls and set 5 as the base plane, etc. For every Monte Carlo simulation a value of 200 was used as the maximum number of runs, considering that models were stable at this number of realizations. In the resulting maps, pixels with no hits were set as null values and therefore no color was assigned to them. For the case of toppling failure, orientations of fractures parallel to foliation were used as basal planes. This set was not considered as a random variable as their orientations are linked to the variations of foliation which are well constrained and homogeneous (chapter 1). No stochastic models of planar and wedge sliding failure were created for the fracture set parallel to foliation as no mean and standard deviation were calculated for these datasets. All resultant probability maps corresponding to the same failure mechanism were overlapped onto an aerial orthophoto of the area in order to evaluate if the location of pixels with a higher probability of failure corresponds with the location of recent rock slope failure deposits or release surfaces. The area of Guggini was used as a test site to check the reliability of the models.

4.6.1 Model validation in selected test areas

The case of plane sliding failure is presented in Figure 4.8. It can be seen that areas of high probability of failure are located at the top of large rock debris cones. These deposits are barely or not vegetated, which is a sign of recent rock slope activity. The results obtained from the stochastic model for plane sliding failure shows that for the area of Guggini 65% of the pixels would not
undergo failure (probability of failure Pf=0) with the modeled conditions. The maximum number of
hits (M) for a single pixel is 577 after 800 runs, this number represents the maximum probability of
failure (Pf) for the area (Pf=0.72, see Table 4.2). The pixels with higher probability of failure are
located in the steeper slopes. Inside of the area surveyed by ground-based photogrammetry (the red
rectangle in Figure 4.8) 42.58% of the pixels have Pf=0. This value drops to only 8.21% for the pixels
inside the digitized release areas of past events (blue polygons in Figure 4.8). 25% of the pixels wit Pf
≥ 0 from the simulation in the area covered by the ground-based elevation model are laying on
former release surfaces. We used parameters from descriptive statistics to characterize the three
samples shown in Figure 4.8 (entire region, area inside of the red rectangle, and release surfaces
marked in blue inside of the red rectangle). Table 4.2 shows the probability of failure by percentiles
for plane sliding. Similar statistics are presented for the cases of wedge sliding and toppling failure
in Tables 4.3 and 4.4.

It can be observed that a considerable part of the pixels lies between values of 0 and 0.2 (Tables
4.2 and 4.5). When pixels with values smaller than 0.2 are not considered in the analysis, the results
fit very well with the release surfaces identified in the ground based photogrammetry elevation
models (Figure 4.9) for the cases of planar and wedge sliding. More than 60% of the pixels with a Pf
> 0.2 inside of the red rectangle fall inside of the identified former release areas. The majority of
pixels falling out of the identified former release areas are located in the surroundings of these
areas (Figure 4.9). In this test area, the highest Pf values correspond to the plane sliding failure and
are located on steep slopes faces (> 75˚) dipping E – NE. A similar spatial disposition but with smaller
Pf values correspond to the results from the wedge sliding model.

For the case of toppling, results show a different spatial disposition than for the other two cases.
Only 30 percent of the pixels with Pf ≥ 0.2 are located in the detected former release areas. However
some of the release areas that were not detected by models of planar and wedge sliding present
relative high values of Pf for toppling (e.g. the release areas located on the upper left quadrant of
the red square in Figure 4.9). Pixels with a Pf ≥ 0.2 for toppling failure are located on steep slopes
(slope angle > 60˚) and facing ESE – SE. As a general observation, in the entire study area it was
observed that rock walls dipping to the SE – S have block toppling as one of the predominant failure
mechanisms. A good example is shown in Figure 4.10 for the area of Walkerschmatt. Rock walls
facing SE – S present relative high values of Pf and fit well with the release areas identified on a
ground based DEM. It is worth to mention that not all S-facing release areas were identified on the
GB-photogrammetric model due to the slope was surveyed from the E. Several south facing slopes
were occluded or difficult to delimit. On the 3D view generated from the ground based DEM shown
on Figure 4.10, the structural control of a toppled block can be observed in detail. The basal plane
follows one of the fractures parallel to foliation (212/18) and the lateral planes are controlled by meso-scale fractures from sets J1 (101/88) and J3 (024/80). Many of the previous failure events in the area shown in Fig. 4.10 had toppling as their mechanism as it can be observed by the characteristics of former release surfaces located on rock walls dipping to the S. In the stochastic simulation of toppling this area has one of the highest Pf values of the observed rock slope (0.35).

<table>
<thead>
<tr>
<th>Percentile</th>
<th>Probability of failure</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Whole slope</td>
</tr>
<tr>
<td>Min.</td>
<td>1.25E-03</td>
</tr>
<tr>
<td>5%</td>
<td>1.25E-03</td>
</tr>
<tr>
<td>10%</td>
<td>1.25E-03</td>
</tr>
<tr>
<td>25% (Q1)</td>
<td>6.25E-03</td>
</tr>
<tr>
<td>50% (Median)</td>
<td>3.13E-02</td>
</tr>
<tr>
<td>75% (Q3)</td>
<td>0.12</td>
</tr>
<tr>
<td>90%</td>
<td>0.25</td>
</tr>
<tr>
<td>95%</td>
<td>0.35</td>
</tr>
<tr>
<td>Max</td>
<td>0.72</td>
</tr>
</tbody>
</table>

*Table 4.2. Summary of statistics for plane sliding on the area of Guggini showed in Figure 4.8, the area covered by GB-photogrammetry (red rectangle in Fig. 4.8), and for the areas corresponding to former release surfaces (blue polygons in Fig. 4.8).*

### 4.6.2 Distributed rockfall probability in the study area

Similar analyses comparing the location of pixels with high Pf with former failure release areas identified with GB-photogrammetry were carried out for other slopes in the study area obtaining similar results. From what was observed, pixels with Pf < 0.2 were discarded and results from the stochastic modeling of all domains with Pf ≥ 0.2 were merged into single maps. The merge was carried out separately for every type of mechanism. Results are presented in Figures 4.11 – 4.13. Finally, the resultant raster datasets from every type of mechanism were combined in a final hazard map. The data was combined such that the raster datasets for each type of mechanism have a similar weight in the final map shown in Figure 4.14.

From the observation of the resultant maps for wedge and planar sliding and for the final map (Figures 4.11, 4.12, and 4.14) it can be observed than the northernmost slope (Sparru) has consistently
the highest Pf values of the whole study area. However, deposits associated with rock slope failures do not reflect well this condition. Debris deposits in the area, present similar volumes than other deposits located to the bottom of other slopes close to Sparru but the deposition processes are different. Only one debris cone is observed which is related to a deeply incised gully. The rest of deposits correspond to rectilinear talus slopes along the bottom of the rock slope parallel to the axis of the Matter Valley. No evidence of large failures is found which implies that the deposits are product of only small size rock slope failures. Lithology is different than the rest of the area too. The lower part of Sparru is formed by chloritic schists with interbedded layers of quartzites with thicknesses up to several meters. Different lithology is associated with different fracture characteristics (e.g. fracture orientation, spacing, persistence, etc), which produce a change in other characteristics of the rock mass (e.g. block size) the change on those characteristics can change the kinematic response of the rock slope. The model proposed in this work, does not consider several of these factors which can produce a change in the results. As the modeling was performed assuming conditions for a specific rock type (i.e.: Randa augengneiss) and only consider the influence of fracture orientation variability and friction angle, the results obtained for Sparru (a slope with different characteristics) probably do not reflect totally its failure conditions and are not similar with the results from the other slopes in the study area. Results from Sparru were not included to quantify statistics of the areal distribution of probability of failure in the area.

<table>
<thead>
<tr>
<th>Percentile</th>
<th>Probability of failure</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Whole slope</td>
</tr>
<tr>
<td>Min.</td>
<td>8.33e-04</td>
</tr>
<tr>
<td>5%</td>
<td>8.33e-04</td>
</tr>
<tr>
<td>10%</td>
<td>8.33e-04</td>
</tr>
<tr>
<td>25% (Q1)</td>
<td>2.5e-03</td>
</tr>
<tr>
<td>50% (Median)</td>
<td>9.17e-03</td>
</tr>
<tr>
<td>75% (Q3)</td>
<td>4.17e-02</td>
</tr>
<tr>
<td>90%</td>
<td>0.13</td>
</tr>
<tr>
<td>95%</td>
<td>0.19</td>
</tr>
<tr>
<td>Max</td>
<td>0.39</td>
</tr>
</tbody>
</table>

Table 4.3. Summary of statistics for the wedge sliding failure on the area of Guggini.
For the rest of the area, a clear difference between the northern and southern parts can be observed. The limit between both spatial domains is located at the valley formed by the Blattbach River (approx. 626800, 111850; Swiss coordinates). Both domains have different area dimensions, the southern domain being ca. 2.5 times larger than the northern domain. In order to make a comparison of the values obtained in both domains in terms of occupied area, results from the northern domain were multiplied by a factor of 2.57 (Table 4.6). As mentioned before, Sparru was not considered for this analysis. Analysis is done with all the Pf values in the considered area for all mechanisms even if for the final hazard map composition pixels with Pf<0.2 were not taken in account.

<table>
<thead>
<tr>
<th>Percentile</th>
<th>Whole slope</th>
<th>Area covered by ground-based model</th>
<th>Release surfaces</th>
</tr>
</thead>
<tbody>
<tr>
<td>Min.</td>
<td>8.33·04</td>
<td>8.33·04</td>
<td>7.17E·02</td>
</tr>
<tr>
<td>5%</td>
<td>5.83·03</td>
<td>1.17E·02</td>
<td>0.13</td>
</tr>
<tr>
<td>10%</td>
<td>1.17·02</td>
<td>2.42E·02</td>
<td>0.16</td>
</tr>
<tr>
<td>25% (Q1)</td>
<td>5.58E·02</td>
<td>8.92E·02</td>
<td>0.21</td>
</tr>
<tr>
<td>50% (Median)</td>
<td>0.13</td>
<td>0.18</td>
<td>0.25</td>
</tr>
<tr>
<td>75% (Q3)</td>
<td>0.21</td>
<td>0.24</td>
<td>0.29</td>
</tr>
<tr>
<td>90%</td>
<td>0.29</td>
<td>0.29</td>
<td>0.32</td>
</tr>
<tr>
<td>95%</td>
<td>0.32</td>
<td>0.31</td>
<td>0.35</td>
</tr>
<tr>
<td>Max</td>
<td>0.43</td>
<td>0.40</td>
<td>0.40</td>
</tr>
</tbody>
</table>

*Table 4.4. Summary of statistics for toppling failure on the area of Guggini.*

From the results described in Table 4.6, plane and toppling failure are more common processes in the northern domain, whereas wedge failure occurs more often on the southern domain. For the case of planar failure, pixels with low Pf (< 0.2) are distributed evenly in both domains (about 1:1) but slightly more abundant in the southern domain. For the rest of classes except the highest class (0.2 ≤ Pf < 0.8) a larger percentage of pixels are located on the northern domain. Pixels from the highest class are only present on the northern domain. Pixels from the higher classes (Pf ≥ 0.4) are preferentially located on rock walls dipping to the E on the southern domain whereas they are located on slopes facing to the N – NE on the northern domain. For the case of wedge failure, pixels from low and intermediate classes (0.1 – 0.5) are predominantly located on the northern domain. Pixels corresponding to higher classes are mostly located on the southern domain. Most of the
pixels located on the higher classes are located on N dipping slopes. Toppling failures are located preferentially in the northern domain for all classes, being the proportion more accentuated on the higher classes (PF > 0.3). However, in general terms, toppling failure has lower probability values than the other two types (maximum Pf = 0.61 on the area of Medji). For the integrative hazard map (Figure 4.14), lower values (PF < 0.2) are located on slopes dipping NW - N – NE. For intermediate values (0.2 < PF < 0.4), pixels are located on SE - E – ENE dipping slopes on the southern domain and on E - NE – NNE facing slopes on the northern domain.

<table>
<thead>
<tr>
<th>Mechanism</th>
<th>Whole slope (Area I)</th>
<th>Area covered by ground-based model (Area II)</th>
<th>Release surfaces (Area III)</th>
<th>% of common pixels between areas II and III</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plane</td>
<td>85</td>
<td>80</td>
<td>50</td>
<td>62</td>
</tr>
<tr>
<td>Wedge</td>
<td>96</td>
<td>90</td>
<td>66</td>
<td>67</td>
</tr>
<tr>
<td>Toppling</td>
<td>72</td>
<td>60</td>
<td>22</td>
<td>30</td>
</tr>
</tbody>
</table>

Table 4.5. Summary of the percentiles occupied by pixels with PF < 0.2 for all failure mechanisms considered. Last column details the percentage of pixels with Pf ≥ 0.2 identified as former release areas.

### 4.6.3 Comparison with failure mode of large rock slope instabilities

The predominant type of failure mode found by the stochastic models often corresponds to observations at currently active large rock slope failures as described in chapter 2. For example, for the case of the area of the Medji rock failure, toppling has been postulated as the main mechanism that controlled the failure in November 2002. The results from the model show that the area of Medji is the one with the highest probability for rock toppling within the whole study area (0.53 – 0.61) on slopes facing to the SE with a similar geometry as the original slope geometry before the 2002 failure. Planar failure, which controlled the failure of one compartment on 2002, presents high Pf (0.57 – 0.78) for the area too, specially for the rock slope steeply dipping to the SE at the crown of the 2002 event. For the area of the current instability at Walkerschmatt, plane failure has been proposed for compartment II (see Chapter 2). The lower limit of the instability which is the lower limit of the compartment II and corresponds to a steep rock wall dipping to the E – ESE shows high values of Pf (0.6 – 0.71) for plane sliding which coincides with the above mentioned considerations. The southern part of the current instability corresponding to the compartment I (see Chapter 2) is controlled by toppling failure which coincides with the results obtained for the area using the stochastic model (PF: 0.26 – 0.33). It is worth to mention that the uppermost part of the instability was not analyzed by the stochastic models as no reliable topographical information was available.
Because of the same condition, the site of Sibulbodme was neither analyzed. Only the lower part of the current instability at Randa could be modeled because of the topographical data availability. The results are consistent with what has been found by Gischig et al. (2010, accepted), being planar sliding the mechanism with the highest \( Pf \) values (0.45 – 0.63) for slopes dipping to the SE (~150°).

<table>
<thead>
<tr>
<th>Pf Class</th>
<th>Plane</th>
<th>Wedge</th>
<th>Toppling</th>
<th>General</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>N</td>
<td>S</td>
<td>N</td>
<td>S</td>
</tr>
<tr>
<td>0.0 – 0.1</td>
<td>44.4</td>
<td>55.6</td>
<td>49.6</td>
<td>50.4</td>
</tr>
<tr>
<td>0.1 – 0.2</td>
<td>49.6</td>
<td>50.4</td>
<td>60.6</td>
<td>39.4</td>
</tr>
<tr>
<td>0.2 – 0.3</td>
<td>57.2</td>
<td>42.8</td>
<td>61.1</td>
<td>38.9</td>
</tr>
<tr>
<td>0.3 – 0.4</td>
<td>61.5</td>
<td>38.5</td>
<td>57.3</td>
<td>42.7</td>
</tr>
<tr>
<td>0.4 – 0.5</td>
<td>60.2</td>
<td>39.8</td>
<td>53.1</td>
<td>46.9</td>
</tr>
<tr>
<td>0.5 – 0.6</td>
<td>55.9</td>
<td>44.1</td>
<td>10.7</td>
<td>89.3</td>
</tr>
<tr>
<td>0.6 – 0.7</td>
<td>55.3</td>
<td>44.7</td>
<td>13.6</td>
<td>86.4</td>
</tr>
<tr>
<td>0.7 – 0.8</td>
<td>64.9</td>
<td>35.1</td>
<td>0</td>
<td>100</td>
</tr>
<tr>
<td>0.8 – 0.9</td>
<td>100</td>
<td>0</td>
<td>100</td>
<td>0</td>
</tr>
</tbody>
</table>

*Table 4.6. Results classified by probability of failure and spatial location (values from Sparru not included).*

### 4.7 Summary

In this paper we describe and apply a new GIS based method for regional rock slope stability analysis. The method is based on probabilistic kinematic analysis (planar sliding, wedge sliding, and single block toppling) and has been designed for small to medium scale rock slope failures.

The method is based on detailed structural characterization of fracture sets and slope orientations from field mapping and high resolution LIDAR digital terrain models with a pixel size of 2.5 meters. From these data sets consistent distributions of fracture set orientations can be derived which are used in kinematic stability analyses using Monte Carlo simulations. The modeling and spatial integration is performed with ArcGIS v. 9.3 using Model Builder and Phyton. This model can also include variability in frictional strength properties and delivers regional distributions of failure probabilities for the kinematic models considered.
The methodology presented here to deal with the uncertainty of spatial variation of fracture orientation applied to kinematic analysis shows a good correspondence with what has been observed during field recognition and was presented in the inventory of rock slope instabilities presented in chapter 2. Location of slopes classified as “stable with rock falls” in the inventory map coincides with the location of the slopes with high probability of failure in the stochastic model. For the areas where current large rock slope instabilities have been identified, the model shows a good fit with the observed mechanisms controlling these instable areas. For all types of failure mechanisms, a higher percentage of the pixels with \( Pf \geq 0 \) are located on the northern domain. These results fit with the observed larger debris deposits and more abundant former release surfaces presented on the inventory of rock slope instabilities presented in Chapter 2.

The results have also been verified with observations of release surfaces on 3D models derived from optical imagery (ground based photogrammetry) and spatial distribution of rock slope failure deposits observed on aerial photographs. The probabilistic GIS modeling results and the observations are consistent and demonstrate that the method has a significant potential for practical applications.

The code can deal with variability of friction angle values too. However, for the present work no detail information of fracture properties controlling friction (e.g.: fracture roughness) was available. More extensive analysis of friction angle values for the area would improve the results. Other fracture properties that control rock slope failure processes, such as persistence or spacing, are not considered in the model.
4.8 References.


Gischig, V., Amann, F., Moore, J.R., Loew, S., Eisenbeiss, H., and Stempfhuber, W., Submitted, Composite rock slope kinematics at the current Randa instability, Switzerland, based on remote sensing and numerical modeling: Engineering Geology.


Jaboyedoff, M., 2002, Matterocking User guide, CREALP (Centre de recherche sur l’environnement alpin).


Sturzenegger, 2009b, Quantifying discontinuity orientation and persistence on high mountain rock slopes and large landslides using terrestrial remote sensing techniques: Natural Hazards and Earth System Sciences, v. 9, p. 21.

SWISSTOPO, 2005, Airborne based LIDAR based 2.5 meters Digital Elevation Model.


Willenberg, and H., 2004, Geologic and kinematic model of a complex landslide in crystalline rock (Randa, Switzerland) [PhD. Thesis thesis]: Zurich, Swiss Federal Institute of Technology.
Figure 4.1. An example of spatial variability of key parameters for kinematic analysis in the area of the Matter Valley. Areas are color coded by main trend of slope orientation. Lines represent meso-scale fracture traces in the rock slope as they are described further in the text.
Figure 4.2. Schematic representation of the geometric relationships of orientation components on a plane. $\alpha = $ dip direction, $\beta = $ dip, $\alpha_s = $ strike. $u_x, u_y, u_z =$ direction cosines of unit vector $u$.

Figure 4.3. Schematic representation of the geometric relationships of orientation components for plane sliding. $\beta = $ dip of discontinuity, $\theta = $ dip of slope, $\gamma = $ angle between dip directions of slope and discontinuity, $U = $ plane of discontinuity, $P =$ slope face.
Figure 4.4. Schematic representation of the geometric relationships of orientation components for wedge sliding. $\beta_1 =$ dip of intersection of discontinuities, $\gamma =$ angle between dip directions of slope and intersections of discontinuities, $U =$ plane of first discontinuity, $V =$ plane of second discontinuity, $P =$ slope face.

Figure 4.5. Schematic representation of the geometric relationships of orientation components for toppling. $\beta_1 =$ dip of intersection of discontinuities, $\theta =$ dip of slope, $\gamma =$ angle between dip directions of slope and dip direction of intersection of discontinuities, $\beta_b =$ dip of basal plane, $U =$ plane of first discontinuity, $V =$ plane of second discontinuity, $B =$ basal plane, $P =$ slope face.
Figure 4.6. Flow diagram of the procedure for evaluating probability of failure for a plane sliding. Equations 1 and 2 are referred to in the text. N = total number of runs, K = count number, M = maximum value for a given raster dataset.
Figure 4.7. Example of GB photogrammetry for Guggini. Left: 3D model highlighting the detected release surfaces.
Right: Stereoplot of the orientation of former release surfaces.

Figure 4.8. Example of results for the area of Guggini for plane sliding failure. Area enclosed on the red square (zoomed to the left to the figure) shows results from GB photogrammetry.
Figure 4.9. Example of the results obtained by the stochastic modeling discarding values smaller than 0.2: a) Plane sliding, b) wedge sliding, and c) toppling. Color legend has been stretched for a better visualization.
Figure 4.10. Example of toppling failure on lateral faces perpendicular to the Matter Valley axis. Left: Results from modeling. Right: 3D model from ground based photogrammetry.
Figure 4.11. Probability of failure by plane sliding on the western flank of the Matter Valley
Figure 4.12. Probability of failure by wedge sliding on the western flank of the Matter Valley
Figure 4.13. Probability of failure by toppling on the western flank of the Matter Valley
Figure 4.14. Integrative probability of failure on the western flank of the Matter Valley
Chapter 5

Summary

The results presented in this work give insights in the control that rock discontinuities exert on the initiation and development of slope instabilities in crystalline rock. The work was carried out in the western flank of the Matter Valley between the localities of Sankt Niklaus and Randa. This area was selected for the study because it offers a unique opportunity to improve the understanding of the relationships between rock discontinuities and rock slope failure phenomena. This section of the Matter valley shows one of the highest topographic gradients in the Alps, a high frequency of landslides, and a high density of settlements and lifelines. The axis of the valley has a N – S orientation in the southern part of the study area and rotates about 15° clockwise in the northern part of the study area (at the intersection with the Blattbach River Valley). Detailed inventories of discontinuity data and landslide phenomena were produced based on field work, analysis of aerial- and ground-based remote sensing imagery, digital elevation models (DEM), and displacements from GPS, InSAR and geodetical monitoring. These inventories were used to establish relationships between rock slope failures and structural characteristics of rock slopes at multiple scales. Additionally, stochastic models representing spatial variability of fracture characteristics were implemented and their results compared with the observations of past failures.

The subsequent sections of this chapter summarize the more relevant outcomes obtained product of the investigations.

5.1 Multi-scale rock mass structure from an integrating field and DEM investigations using geographic information systems

Structural characterization for the western flank of the Matter Valley has been carried out for the present work at multiple scales. Characterization has been focused in discontinuity orientation. Discontinuities have being divided on penetrative and non-penetrative. Penetrative discontinuities are represented only by foliation in the area. Non-penetrative discontinuities are rock fractures of different types: joints, faults, shear zones and fracture zones. Fractures have been sub-divided arbitrarily on meso- and large-scale discontinuities based on their length. Spatial variation of their orientation has been observed through a detailed analysis of data.
Smooth changes in orientation of foliation could be observed from S to N and from bottom to top of the rock slopes in the study area. Continuous surfaces of dip and dip direction values were constructed interpolating measurements from field data using disjunctive kriging. Dip direction values vary from S to NW (180° - 305°) in the area and dip angle values from gentle to slightly steep values (12° - 35°). Two sets of non-penetrative discontinuities follow orientations parallel to foliation. Based on the interpolated surfaces, conformity between topography and foliation was evaluated. Results show that slopes in the area are predominantly orthoclinal (49%) and anacinal (47%) with steepened escarpments. Overdip slopes occupy only 4% of the study area but are important in terms of slope stability. They mainly occur on slopes dipping to the S and SW. Such overdip slopes are not uniformly distributed and mainly occur in the southern part of the area in northern flanks of side valleys below 2000 m asl.

Meso-scale fractures were defined as rock breaks where only minor tectonic displacement has been found and which possess persistence values in the order of meters to tens of meters. A new GIS based approach has been developed to extract large numbers of meso-scale fracture orientations from high resolution digital elevation models (DEM). Selection of the areas was carefully done based on observations collected during field work and interpretation of aerial images and field photographs. The approach is based on automated sampling of surface orientations from manually selected outcrops. The data was corrected from a systematic bias introduced by the vertical projection of sampled steep planes. The reliability of these computer based data sets has been tested by comparison with the field data (2110 fracture measurements) using classic statistics (ANOVA tests). For the analysis, the study area was subdivided into 10 spatial domains. Fisher distributions were used to observe clusters and group the data into sets. Five systematic meso-scale sets were identified in most domains dipping preferentially to the N, E and NE. Spatial variation of fracture orientations for all discontinuity sets was observed from S to N of the study area. Several types of probability distributions for dip and dip direction of all sets were tested in order to find the best-fit distribution to the measured samples in order to identify the best models for orientation variations of meso-scale discontinuities in all structural domains. Kolmogorov-Smirnov and Anderson-Darling tests were used to assess the goodness of fit of the proposed distributions. Normal, Weibull and Gamma distributions obtained the highest scores. Based on the changes on fracture orientation, the study area can be divided in two regions with significant differences whose limit is located at the valley of the Blattbach River.

Large-scale discontinuities corresponding to brittle faults, fracture zones and brittle-ductile shear zones were identified in the study area based on field investigation. Four sets were found in the area through field data collection (226 measurements) and DEM analysis of 350 lineaments
traces using a Matlab code based on multiple point least squares regression. DEM-derived fault orientations show a significantly higher scatter than field-based data.

5.2 Rockslide susceptibility assessment derived from integrated field and remote sensing investigations.

An inventory map of rock slope instability phenomena was produced covering the whole study area. Three main units related with slope stability were used: deposits, past events and current instabilities. These classes were subdivided based on specific characteristics for each unit. Slopes that were not classified in one of these three units were considered as stables. An additional classification of the stable slopes was done based on the presence or not of small size rockfalls. The regional comparison of the density and magnitude of release areas and deposits shows that the study area can be subdivided in northern and southern regions separated by the Blattbach River. These two domains also indicate a different geomorphological evolution expressed by differences in the frequency, depth and width of gullies eroded into the steep sidewalls of the western Matter valley. In the southern domain, many (lateglacial) release surfaces have been found, mostly located on the front walls of the main valley. In this domain, magnitudes of the deposits are comparatively small. In the northern domain, magnitudes of rock fall deposits are systematically larger than in the south, being in the order of several tens of thousands of square meters. Also larger debris flow cones exist in the northern domain compared to the ones in the southern domain. In this domain most (postglacial) slope failures with genetically correlated deposits are located in the steep walls of the main valley slopes and a few in the southwestern-facing flanks of the lateral valleys.

Intermediate and large current rock slope instabilities are widespread on the study area and are not specifically concentrated in the southern or northern domain. The instabilities occur at different heights (from ~1800 to ~2400 m asl), mainly on steep slopes facing to the S or SE.

Three of the larger active instabilities in the Matter Valley were studied in more detail: Walkerschmatt, Medji and Sibulbodme. Regional fracture sets, both meso- and large-scale, found on the structural characterization, play an important role in these instabilities.

It can be observed that slope geometry has a relationship with the type of mechanism controlling failure. E-NE facing slopes are prone to fail by sliding (northern compartment Walkerschmatt, Sibulbodme), whereas SE-S facing slopes are more susceptible to fail by toppling (southern compartment Walkerschmatt, Medji rockslide 2002). Sliding occurs along pre-existent, moderately steep fractures (set F2 for the cases above mentioned) and occurred on either moderately steep or steep slopes. Toppling is controlled by steep fractures (mainly set J4, intersected
with set $J_1$ as lateral boundaries) and foliation parallel fractures (as basal planes). Toppling occurs on steep rock faces. In most cases, internal deformations of the rock masses are controlled by steep meso-scale fracture sets, and large discontinuities (faults) control the boundaries of the instabilities (e.g. Sibulbodme, Medji, Randa).

### 5.3 GIS-based stochastic methods for kinematic stability analysis of rock slopes

Small size rock fall spatial distribution has been assessed by applying physically-based methods for three classic failure mechanisms: planar sliding, wedge sliding and toppling. Monte Carlo simulations of fracture distributions have been applied in a geographic information system environment using commercial software (ArcGIS) to quantify probability of failure ($P_f$) based on principles of kinematic analysis. Normal distribution parameters (mean and standard deviation) for orientation of meso-scale discontinuities as derived from the structural characterization of the area were used. The models also include variability in frictional strength properties and delivers regional distributions of failure probabilities for the considered failure mechanisms.

The results obtained of the application of the methodology presented here to deal with the uncertainty of spatial variation of fracture orientation applied to kinematic analysis shows a good correspondence with what has been observed during field recognition and was presented in the inventory of rock slope instabilities. Location of stable slopes with presence of small size rockfalls in the inventory map coincides with the location of areas of high $P_f$ values in the models. For areas occupied by current large rock slope instabilities, the types of mechanisms with the higher values from the models correspond to the observed mechanisms controlling these instable areas. For all types of failure mechanisms, a higher percentage of the pixels with $P_f \geq 0$ are located on the northern domain. These results fit with the observed larger debris deposits and more abundant former release surfaces observed in the northern half of the area on the inventory map of instabilities.

The results have also been verified with observations of release surfaces on 3D models derived from optical imagery (ground based photogrammetry) and spatial distribution of rock slope failure deposits observed on aerial photographs. The probabilistic GIS modeling results are consistent with the field observations and demonstrate that the method has a significant potential for practical applications.
Curriculum Vitae

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