Quaternary subglacial processes in Switzerland: geomorphology of the Plateau and seismic stratigraphy of Western Lake Geneva

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Abstract

Les processus sous-glaciaires quaternaires dans l'avant-pays suisse sont étudiés ici au moyen de la géomorphologie et de la sismo-stratigraphie. L'étude géomorphologique commence par une modélisation du substat rocheux de la Suisse occidentale mettant en évidence un réseau anastomosé de vallées-tunnel sur le plateau. Puis, sur la base de modèles numériques de terrain, les formes glaciaires héritées des dernières glaciations (drumlins rocheux, drumlins, ribbed moraines, eskers et moraines) sont cartographiées et analysées. Ceci constitue la première mention de ribbed moraines sous la calotte alpine. La sismo-stratigraphie du Petit-Lac (partie occidentale du Léman) est basée sur de nombreux profils sismiques haute-résolution. 12 unités sont différenciées. La plupart d'entre elles sont des tills déposés sous le glacier flottant, à mesure que l'épaisseur de glace diminuait. De nombreuses oscillations du front glaciaire sont reconnues. L'étude révèle également la présence d'eskers associés à un éventail pro-glaciaire.

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Quaternary Subglacial Processes in Switzerland: Geomorphology of the Plateau and Seismic Stratigraphy of Western Lake Geneva

THESE

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GENEVE
The subglacial environment is difficult to reach and its investigation relies on indirect observations, such as the study of subglacial landforms and sediments in formerly glaciated areas. The goal of the present work is to shed new light onto past subglacial processes in the Swiss Alpine foreland, which were poorly studied until now. For this purpose, two complementary methods have been used: geomorphology and seismic stratigraphy. The analysis of digital elevation models (DEMs) evidences the landforms inherited from the glaciations, whereas the seismic stratigraphy of Western Lake Geneva highlights the subglacial sedimentary sequence accumulated in glacial valleys.

Geomorphology

The first part of this work investigates the geomorphology of bedrock and glacial landforms (drumlins, ribbed moraines, eskers and moraines). The study of the bedrock in Western Switzerland is based on the compilation of pre-existing data completed by new data from seismic reflection. The study of the alpine foreland landforms is based on two digital elevation models (DEM): the national Swiss DEM (DHM25), covering the entire Swiss territory, and the high-resolution LiDAR DEM covering the cantons of Geneva and Vaud.

Several topographic attributes were derived from the DEMs: hillshading, slope, profile curvature, openness and low-cut (high-pass) filtered elevation. The last attribute is particularly useful because it suppresses the low-frequency regional topographic variations and preserves the high-frequency local topographic variations, including glacial landforms. From these attributes, landforms have been digitally mapped. This digital format allowed the extraction of statistics regarding their dimensions and distribution.

The Alpine range presents a dendritic network of V-shaped valleys cutting the bedrock down to several hundreds of meters below present sea level. In these valleys, meltwater flowed in a single channel at the bottom of the main valley. On the northern Alpine foreland plateau, these valleys evolve into tunnel-channels. The latter are shallower, deeper, with an undulated or concave longitudinal profile, and form in places an anastomosed network. This morphological change between the Alpine range and the foreland reflects the transition from Alpine to piedmont glaciers. These valleys were eroded mainly by subglacial meltwater circulation. In the Alpine range, the meltwater path was influenced by the important variations in elevation, whereas in the foreland, this path depended mainly on the hydrostatic pressure and, therefore, on the ice-surface topography.

Between these valleys, the foreland presents several subglacial landforms: rock drumlins, drumlins, ribbed moraines and eskers. Rock drumlins, recognised between Lausanne and Soleure, result from the differential glacial erosion of the Molasse marls and sandstones by the Rhône glacier. Most of them are oriented parallel to the strike of the Molasse beds and have an asymmetric cross-section due to the Molasse dip. Other rock drumlins of smaller dimensions are perpendicular to the Molasse beds orientation.

1237 drumlins were mapped, grouped in 22 fields related to the Jura ice cap and to the Arve, Rhône, Aar, Reuss, Linth and Rhine glaciers. These drumlins are made of eroded glacial sediments and occur near the external moraines of the Würm maximum, but not in contact with them. The average dimensions vary between fields, with width ranging from 65 to 245 m, length/width ratio from 2 to 3 and height/width ratio from 0.07 to 0.12. In six fields, 273 ribbed moraines were recognized, representing the first record of these landforms in the Alpine ice cap area. Ribbed moraine drumlinisation and the progressive spatial transition between these two forms suggests a synchronous formation by a similar physical process. Sedimentological evidences indicate an erosional process but do not determine the erosive agent (ice or water).

Three esker systems have been identified in the Vaud canton. 1) The Bois-de-Chênes, north of Nyon, represents a braided esker system made of numerous small ridges formed by meltwater coming from the Jura range. 2)
The Ballens complex, in the Bière area, is a marginal fluvio-glacial complex formed by the arrival of subglacial meltwater into an ice-marginal lake. 3) A very well preserved esker lies in the Joux Valley, formed below the Jura ice cap. These eskers point to important meltwater circulation below the glacier.

LiDAR data also showed numerous recessional moraines. The most spectacular is that of Montosset, in the Bière area, which is 15 km long and presents numerous closely spaced crests indicating oscillation of the ice front during the retreat.

This work is the first global study of drumlins at the scale of Switzerland and the first record of ribbed moraines under the former Alpine and Jura ice caps. The association of subglacial landforms such as tunnel-valleys, rock drumlins, drumlins, ribbed moraines and eskers had only been reported until now under larger ice sheets. Their presence under the Alpine and Jura ice caps indicates similar sub-glacial processes, despite the smaller size of these ice bodies. Drumlins, ribbed moraines and eskers indicate that the ice base was above the melting point during their formation.

Western Lake Geneva seismic stratigraphy

The seismic stratigraphy of the Petit-Lac (Western part of Lake Geneva) is based on a dense grid of high-resolution, multi-channel, airgun seismic profiles. The Petit-Lac is located in a valley cut into the Molasse bedrock, which shallows up and narrows towards the southwest. This valley is filled with a thick sequence (up to 220 m) of glacial, glacio-lacustrine and lacustrine sediments, divided in 14 seismic units (U1 to U14).

The glacial sequence is complex, with numerous erosion surfaces and rapid lateral changes in seismic facies. This complexity points out to a highly dynamic sedimentary system with interaction of ice and water flows. U1 represents remnants of glacial deposits older than the last glacial cycle, preserved in the deepest part of the lake and in secondary bedrock valleys. U2 represents gravel and sands deposited by meltwater circulation at the bottom of the glacial valley. U3 is a thick, stratified unit marking the beginning of the deglaciation, when the Rhône glacier became thinner and buoyant and allowed the formation of a subglacial lake. Younger glacial units (U4, U5, U7, U9, U11) are acoustically chaotic sediments deposited subglacially under the water table, while the glacier was thinning. These glacial units are bounded by synform erosion surfaces corresponding to small readvances of the glacier. These erosion surfaces are well defined in the northern part of the Petit-Lac and merge towards the south, where younger erosion surfaces truncate older ones.

The transition from the glacial to the glacio-lacustrine environment was progressive and started with the apparition of a marginal esker-fan system (U6). Esker formation was followed by a small advance-retreat cycle leading to the deposition of U7. Then the ice front receded and stratified sediments were deposited in a glacio-lacustrine environment (U8, U10, U12). This retreat was punctuated by two readvances – Coppet (U9) and Nyon (U11) – producing large push moraines and proglacial debris flows. The presence of push-moraines clearly indicates a steep ice front and oscillations of the glacier front during the deglaciation. Finally, a lacustrine environment took place.

Regarding eskers, DEM analysis and seismic stratigraphy in Lake Geneva are complementary methods to study subglacial processes in Switzerland. Their combination has revealed the path of eskers in the Petit-Lac and their absence on Lake Geneva shores.
L'environnement sous-glaciaire est difficile d'accès et son étude se base par conséquent sur des observations indirectes, notamment l'étude des morphologies et sédiments sous-glaciaires dans les zones autrefois englacées. La présente thèse a pour but de mieux comprendre les processus sous-glaciaires quaternaires dans l'avant-pays alpin suisse, peu étudiés jusqu'à présent. Dans ce but, deux méthodes complémentaires ont été utilisées : la géomorphologie et la sismo-stratigraphie. L'analyse des modèles numériques de terrain (MNA) met en évidence les morphologies héritées des glaciers tandis que la sismo-stratigraphie de la partie occidentale du Léman révèle la séquence sédimentaire sous-glaciaire accumulé dans les vallées glaciaires.

Géomorphologie

La première partie de ce travail s'intéresse à la géomorphologie du substrat rocheux et aux formes glaciaires (drumlins, ribbed moraines, eskers et moraines). L'étude morphologique du substrat rocheux en Suisse occidentale est basée sur la compilation de données préexistantes complétées par de nouvelles données issues de la sismique réflexion. L'étude du relief de l'avant-pays alpin est issue de l'analyse de deux MNA: le modèle national suisse à 25 m (DHM25) couvrant toute la Suisse et le modèle haute-résolution LiDAR (light detection and ranging) couvrant les cantons de Genève et Vaud.

Plusieurs attributs de surface ont été dérivés des MNA: ombraje, pente, courbure, « openness » et filtrage coupe-bas des fréquences. Ce dernier est particulièrement utile car il supprime les variations topographiques régionales de grande longueur d’onde et préserve les variations locales de faible longueur d’onde incluant les morphologies glaciaires. A partir de ces attributs, les morphologies glaciaires ont été cartographiées de manière vectorielle. Ce format digital a permis d'extraire des statistiques concernant leur distribution et leur morphologie.

La chaîne alpine présente un réseau dendritique de vallées en V qui entaillent le substrat jusqu'à plusieurs centaines de mètres sous le niveau de la mer. Dans ces vallées, l'eau de fonte circulait dans un chenal unique situé au fond de la vallée principale. Sur le plateau, ces vallées alpines évoluent en vallées-tunnels moins profondes, plus larges, au profil longitudinal ondulé ou concave, formant par endroits un réseau anastomosé. Ce changement de morphologie entre la chaîne alpine et le plateau reflète la transition entre des glaciers alpins et des glaciers de piedmont. Ces vallées ont été érodées principalement par les circulations d'eau de fonte à la base du glacier. Dans la partie alpine, le relief prononcé du substrat dictait le chemin emprunté par l'eau de fonte, tandis que sur le plateau ce chemin dépendait principalement de la pression hydrostatique liée à la topographie de la surface du glacier.

Entre ces vallées, le plateau présente diverses morphologies sous-glaciaires : drumlins rocheux, drumlins, ribbed moraines et eskers. Les drumlins rocheux, reconnus entre Lausanne et Soleure, résultent de l'érosion glaciaire différentielle des marnes et grès molassiques par le glacier du Rhône. Ils sont généralement orientés parallèlement à la direction des bancs de Molasse et présentent une asymétrie transverse liée au pendage de la Molasse. D'autres drumlins rocheux, plus petits, se sont formés perpendiculairement à l'orientation des bancs de Molasse.

1237 drumlins non-rocheux ont été reconnus, regroupés dans 22 champs liés à la calotte jurassienne et aux glaciers de l'Arve, du Rhône, de l'Aar, de la Reuss, de la Linth et du Rhin. Les drumlins sont formés de sédiments glaciaires divers et se situent proches des moraines externes du maximum würmien, mais pas directement contre celles-ci. D'un champ à l'autre, les dimensions moyennes varient de 65 à 245 m pour la largeur, 2 à 3 pour le rapport longueur/largeur et 0.07 à 0.12 pour le rapport hauteur/largeur. 273 ribbed moraines (ou Rogen moraines) ont été reconnues dans six champs, constituant la toute première mention de ces formes dans le périmètre de la calotte glaciaire alpine. La drumlinisation des ribbed moraines et la transition spatiale progressive entre les deux formes suggèrent une formation synchrone par un processus physique
similar. Les évidences sédimentologiques indiquent que ce processus est érosif mais ne permettent pas d’attribuer cette érosion à la glace ou à la circulation sous-glaciaire d’eau de fonte.

Trois systèmes fluvio-glaciaires de type esker ont été identifiés dans le canton de Vaud. 1) Le Bois de Chênes, au nord de Nyon, représente un système d’esker en tresses constitué de nombreuses crêtes de petites dimensions, formées par un apport d’eau du Jura. 2) Le complexe de Ballens, dans la région de Bière, est un système fluvio-glaciaire marginal formé par l’arrivée de circulations sous-glaciaire dans un lac péri-glaciaire. 3) Un esker très bien conservé formé sous la calotte jurassienne a été reconnu dans la Vallée de Joux. Ces eskers attestent des importantes circulations d’eau de fonte à la base du glacier.

Les données LiDAR ont également permis d’identifier de nombreuses moraines de retrait. Celle de Montosset, dans la région de Bière, se suit sur plus de 15 km et présente de nombreuses crêtes successives attestant des oscillations du front du glacier durant le retrait.

Ce travail constitue la première étude des drumlins à l’échelle de la Suisse et la première reconnaissance de ribbed moraines dans le périmètre des glaciations alpines. L’association des formes sous-glaciaires telles que vallées-tunnels, drumlins rocheux, drumlins, ribbed moraines et eskers n’avait été jusqu’à maintenant reportée que sous des calottes glaciaires plus importantes. Leur présence indique que les calottes alpine et jurassienne présentaient un système sous-glaciaire similaire, malgré leur dimension inférieure. Les drumlins, les ribbed moraines et les eskers indiquent que la base de la glace était au-dessus du point de fusion durant leur formation.

Sismo-stratigraphie du Petit-Lac

La sismo-stratigraphie du Petit-Lac (partie occidentale du Léman) est basée sur une grille serrée de profils sismiques haute-résolution multi-canaux.

Le Petit-Lac est situé dans une vallée entaillée dans le substratum molassique, qui rétrécit et devient moins profond vers le sud-ouest. Cette vallée est remplie avec une épaisse séquence (jusqu’à 220 m) de sédiments glaciaires, glacio-lacustres et lacustres divisée en 14 unités sismiques (U1 à U14).

La séquence glaciaire du Petit-Lac est complexe, avec de nombreuses surfaces d’érosion et des variations latérales rapides de faciès sismique. Cette complexité traduit un système sédimentaire très dynamique avec une interaction entre les processus liés à la glace et à l’eau. U1 représente les restes de dépôts glaciaires plus anciens que le dernier cycle glaciaire, préservés dans le fond de la vallée et dans des vallées latérales secondaires. U2 représente des graviers et des sables déposés par la circulation d’eau de fonte au fond de la vallée. U3 est caractérisée par des strates épaisse déposées dans un lac sous-glaciaire au début de la déglaciation, lorsque l’épaisseur de glace devint assez fine pour permettre sa flottaison. Les unités glaciaires suivantes (U4, U5, U7, U9, U11) ont un faciès sismique chaotic à semi-transparent et ont été déposées sous le glacier flottant. L’aggradation des unités traduit l’amincissement progressif de la glace, tandis que les réflecteurs les séparant sont des surfaces d’érosion liées à des petites réavancées du glacier. Ces surfaces d’érosion remontent vers les bords du lac et se rejoignent vers le sud-ouest, où les surfaces d’érosion plus jeunes tronquent les plus anciennes.

La première indication d’un lac pro-glaciaire et d’un environnement de type fjord est la présence d’un esker associé à un éventail sous-lacuste glacio-marginal (U6). La formation de cet esker a été suivi d’un cycle mineur de réavancée-retrait conduisant au dépôt marginaux de U7. Ensuite, le front glaciaire a reculé, laissant place à des dépôts glacio-lacustres (U8, U10, U12). Ce retrait a été ponctué de deux réavancées – Coppet (U9) et Nyon (U11) – qui ont produit d’importantes push-moraines d’où s’écoulaient des debris-flows. Ces push-moraines indiquent un glacier au front abrupt. Finalement, un environnement lacustre similaire à celui actuel s’est mis en place.

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1.1 Aims of the study

The subglacial environment involves complex interactions between ice, water, and sediments. If some subglacial landforms are fairly well understood (e.g., eskers), glacial geomorphology is still a matter of debate, mainly concerning the formation of drumlins and ribbed moraine. This poor understanding results from the difficulty to access this environment for a direct investigation and the absence of such subglacial landforms at the front of glaciers receding today, suggesting different physical conditions below the past ice sheets. Investigation of past subglacial processes is therefore based mainly on indirect observations, through the study of subglacial sediments and landforms. Many of these studies have been conducted in formerly glaciated areas like North America, Fenno-Scandinavia or Ireland. In the Alpine area, however, little attention has been paid to past subglacial conditions. The present study aims therefore to investigate these subglacial conditions in this particular area, focusing on Switzerland.

For this purpose, two methods have been used: geomorphology (part I) and seismic stratigraphy (part II). Part I concerns the distribution and morphology of subglacial landforms in the Swiss Alpine foreland and adjacent areas, through the analysis of digital elevation models (DEMs). This geomorphological analysis is particularly detailed in Western Switzerland, thanks to the availability of high-resolution LiDAR data. Part II focuses on the Quaternary infill of western Lake Geneva (Petit-Lac), based on high-resolution, multi-channel seismic reflection. These two methods provide complementary informations about subglacial processes between and within valleys and permit to link both environments. For the Petit-Lac area, the association of these two methods should allow to determine if the eskers visible on seismic reflection in the Petit-Lac extend outside the lake.

1.2 Alpine glaciations

1.2.1 Würm Glacial Maximum paleogeography

During the last glacial maximum (LGM, ~ 21 kyr BP), ice covered a large part of the Northern hemisphere (Fig. 1-1). In comparison to the large North-American or Eurasian ice sheets, the Alps presented a relatively small ice mass composed of numerous coalescing valley glaciers, partly individualized by high bedrock crests emerging as nunataks. In Switzerland, the timing of the last large glaciation is contentious: it may have occurred earlier than the LGM in the western part of the Alps. This maximum occurred during the last glacial period – Würm – and is here referred to as the Würm Glacial Maximum (WGM) to avoid confusion with the global LGM. During the WGM, spatial variation in snow accumulation in the innermost Alps led to the formation of small ice domes evolving laterally into ice streams guided by the subglacial valleys (Florineth & Schluechter, 1998; Kelly et al., 2004). When reaching the Alpine foreland, the valley glaciers spread into large piedmont glaciers. This ice mass is traditionally called the Alpine ice sheet, although the term icefield would be more adequate given the very strong influence of the underlying topography upon the ice flow (Benn & Evans, 1998).

The only published map of the WGM ice surface in Switzerland is that of Jäckly (1970). This work is still valid today, even if slight modifications would be necessary to account for the recently discovered inner-alpine ice domes and the small glaciers of the “unglaciated” Napf area (Schluechter, 2004). The Alpine ice sheet reconstruction has been completed by van Husen (1987) and Keller and Krayss (2005) for the eastern Alps, De Beaulieu et al. (1991) and Guiter et al. (2006) for the southwestern Alps and Campy & Arn (1991) for the Jura ice cap (Fig. 1-2).

During the WGM, two major glaciers were coming down from the Alps into the northern alpine foreland: the Rhone glacier to the west, draining the southwestern
Swiss Alps, and the Rhine glacier to the east, draining the northeastern Swiss Alps. Between them, smaller glaciers reached the foreland: the Aar, Reuss and Linth glaciers. When coming out from the Alps, the Rhine glacier formed a large Piedmont glacier centered on Lake Constance. To the west, this ice mass was partly connected to the smaller Linth and Reuss glaciers.

Because the northern Alpine foreland pinches out towards the SW, the situation was different for the Rhone glacier, which could not form a large piedmont glacier. Blocked by the Jura range, it split into two lobes flowing NE and SW. The northeastern lobe was joined by the Aar glacier in the region of Bern and flowed until Wangen a.A., beyond the town of Solothurn. The southwestern lobe flowed into the Lake Geneva basin, passed over Geneva and ended in the Lyon area (France). For a more complete introduction to the Rhone glacier, see Badoux (1995).

Along the Jura range, the Rhone glacier was in contact with the Jura ice cap (Fig. 1-2), which culminated at 2000 m and flowed mainly towards the SE where it joined the Alpine glaciers (Campy, 1992). At the end of the Würm glaciation, when the receding Rhone glacier left the Jura border, the Jura glaciers could readvance down to the Jura foothill, an episode known as “récurrence des glaciers jurassiens” (Schartd, 1898; Aeberhardt, 1901). According to Arn (1984), the Rhone glacier liberated the Jura ice cap time-transgressively and the Jura glaciers readvance had a different age depending on the area. The same author showed that the maximal extent of this readvance is always associated to kame terraces built between the Jura and Rhone ice fronts.

In this study, the WGM ice extent for the Alps (Florineth & Schluechter, 1998) and the Jura (Campy, 1992) have been digitized in the shapefile WGM _ Alps _ Jura.shp, present on the accompanying CD-ROM (Appendix F).

1.2.2 Chronology

In the Alps, the Pleistocene has been classically subdivided after the scale of Penck and Brückner (1901-1909) into four glaciations: Günz, Mindel, Riss and Würm. Today, this terrestrial scale tends to be replaced by the continuous marine isotope record (Wright, 2000), which shows that glacial cycles were much more numerous (up to 20) and that higher frequency climatic variations were superimposed on them. Today, the relation between the local and fragmentary alpine record and the continuous marine record is controversial, as interpretation of the ice
extent variation through the Middle and Late Pleistocene differs greatly between regions and authors.

On the northern alpine foreland, two main morainic systems exist. The most external corresponds to the most extensive glaciation (MEG, formerly called Riss) and the most internal to the Würmian glacial maximum (WGM, traditionally called Würm). The MEG is generally considered to be the penultimate glaciation (ca 190-130 ky BP), but it could be older: 230-145 ky BP (Schlüchter, 2004) or even older than four glacial cycles (Schlüchter & Kelly, 2000).

As far as the WGM is concerned, its datation and duration are also a matter of debate. On the Swiss plateau, the WGM is thought to correspond to the global last glacial maximum (LGM), situated around 21'000 yr cal. BP with a duration of 3000 to 4000 years (van Husen, 1997; Preusser & Schlüchter, 2004) and corresponds to marine isotope stage 2 (MIS 2). Major glacier retreat took place between 18 and 16 ka cal. BP, punctuated by nine readvances forming sets of fronto-lateral moraines in the alpine foreland (van Husen, 1997). The Alpine foreland was almost ice-free after 17 ka cal. BP and after 16 ka cal. BP, the glaciers receded from major valleys, still interrupted by short-term readvances, for finally splitting into small disconnected valley glaciers. During the Younger Dryas ('Egesen' stage c.a. 12 ka cal. BP), the ice readvance was not important enough to reach the major valleys. In the French Alps, $^{14}$C dating and palynological correlations suggest an earlier WGM, reached during MIS 4 (Early Würm, ~ 60 kyr BP) (Guiter et al., 2006). The discrepancy of the datations in the western and eastern Alps suggests an asynchronous WGM, related to a climatic gradient from oceanic to continental (Brun & Hannss, 1998; Guiter et al., 2006).

Regarding the Jura ice cap, the study of paleo-lakes from the Jura northwestern margin reveals that the WGM lasted from 25,500 to 22,000 cal. yr BC and the glacial retreat from 22,000 to 17,000 cal. yr BC, punctuated by five episodes of stabilization (Buoncristiani & Campy, 2004).
1.2.3 Subglacial thermal regime

Subglacial processes depend on the mechanical properties of the glacier substrate (soft-bedded / hard-bedded) and on the thermal regime at the ice-bed interface (melting/warm-based or non-melting/cold-based) (Clarke, 2005). Subglacial erosion (abrasion) is much stronger below a warm-based glacier than below a cold-based glacier because of the negligible rate of basal sliding of the latter.

Today, most Alpine glaciers are warm-based but cold-based ice exists at high elevations (>3600 m) and where ice is thin or oriented towards north or west (Haeberli, 1975). Regarding the subglacial thermal regime during the former glaciations, authors disagree.

In overdeepened Alpine valleys, glacial sediments of the penultimate glaciation (Riss or Early Würm) lie directly on the Molasse bedrock. This contact implies a powerful erosion of all the glacial material deposited during previous glaciations and thus warm-based conditions (van der Meer, 1982).

For the WGM, Haeberli and Schlüchter (1987) suggested a large predominance of cold-based conditions, based on theoretical considerations. The preservation of Rissian till below the Würmian till would also indicate less erosion during the WGM than during previous glacial cycles, suggesting cold-based conditions (van der Meer, 1982). On the other hand, deposition of lodgement till itself (van der Meer, 1982) and the presence of overdeepened V-shaped valleys, eskers and jökulhaup deposits (Pugin, 1989) are indicative of warm-based conditions. WGM subglacial thermal regime was therefore either warm-based or polythermal.
Part I

Glacial Landforms
of the Swiss Plateau
2.1 Introduction

High-resolution digital elevation models (DEMs) from satellite or airborne data become increasingly available and constitute a primary information source for glacial research, because they provide both an overview of large areas and morphological details. DEM study is generally associated to a GIS (Geographic Information System) environment, allowing the integration of different datasets. A good example of such integration is the work of Benz (2003) on the WGM Rhein-Linth glacier, where combination of ice surface, bedrock surface and present topography allowed the calculation of ice thickness, shear stress, ice velocity and subglacial drainage.

The first part of this work investigates past subglacial processes on the Swiss northern Alpine foreland with particular attention on its western part, based on the bedrock large-scale erosion and on the morphology and spatial distribution of glacial landforms.

Chapter 2
Introduction to Part I

2.2 Glacial landforms

Most of the former studies on Alpine glaciations focused on fronto-lateral moraines, with little interest on streamlined bedforms and eskers. Characteristics and mode of formation of the latter are presented below.

2.2.1 Streamlined bedforms

Definitions

The term 'streamlined bedforms' refers to three types of subglacial landforms: drumlins, megaflutings and ribbed moraines. Drumlins are elongated, oval or tear-shaped hills formed subglacially, with blunted end dipping steeply up-ice (stoss side) and a pointed end dipping gently down-ice (lee side) (Menzies, 1979). Variations of this typical shape are common and include spindle drumlins (long and narrow, pointed at both ends), parabolic drumlins (widening and tapering down-flow, sometimes similar to barchan dunes) and transverse asymmetrical drumlins (coalescent drumlins aligned obliquely to the flow) (Shaw, 1983). Drumlin dimensions are typically 100-2000 m long and 5-50 m high with a variable elongation ratio (length/width). When this ratio is very high (> 10), they are called megaflutings (or megaflutes, mega-scale flutings of flute-ridges), all these names referring to their similarity with the small flutings found at the front of modern glaciers. Because these two bedforms form a continuum (Fig. 2-1), the distinction between drumlins and megaflutings is arbitrary (Rose, 1987). They occur in large groups named drumlin fields generally located close to the outer margins of past ice sheets, often in low-lying piedmont areas where ice flow was diverging (Menzies, 1979 and

The data presented here are available as shapefiles on the attached CD-ROM (Appendix F), together with four GIS projects allowing easy navigation through the maps. Processing methods are explained with more details in Appendix A.
Large drumlin fields exist in many formerly glaciated areas such as North America, Ireland or Scandinavia.

Drumlins are generally found within areas of thick glacial deposits (Hoppe, 1959). They can also occur in areas of scoured bedrock with scattered patches of thin till deposits (Menzies, 1979) or consist entirely of bedrock. Glacial deposits lithology and structure is highly variable: sediments may be till or glacio-fluvial sediments, massive or stratified, with or without glacio-tectonic structures, conformable or discordant with the drumlin surface. Some of the drumlins formed mainly of glacial deposits have a resistant core formed of bedrock or competent till (Fig. 2.2).

The physical environment of drumlin formation was investigated by Patterson and Hooke (1995), based on a review of 95 published descriptions of drumlin fields worldwide. They concluded that drumlins occur: 1) in the ablation area, where ice is affected by compressive longitudinal strain rate, often associated to an extensive transverse strain for fan-shaped drumlin fields, 2) under relatively thin ice, not directly against the ice margin but slightly up-ice, 3) under high pore-water pressure, due to either a frozen margin blocking the water flow or a water-terminating ice front. On the other hand, no relation seems to exist with the lithology of drumlin-forming sediment or with the lithology or large-scale topography of the bed.

Drumlins must be distinguished from crag-and-tails, which consist of a prominent bedrock obstacle leading to the deposition or preservation of sediments behind it. Rock drumlins (also called tadpole rocks, whalebacks or rock ridges) are erosional drumlins consisting exclusively of bedrock. The latter bedforms form under a thick ice cover of at least several hundreds of meters (Evans, 1996). They must not be confounded with roches moutonnées, smaller bedrock bumps or hills with smooth up-ice faces and quarried down-ice faces, formed under a thinner ice cover (Benn & Evans, 1998).

Ribbed moraines (also called Rogen moraines) consist in an assemblage of numerous, sub-parallel, closely and evenly spaced ridges of similar height, oriented transverse to former ice flow. They are generally 5-50 m high, 100-700 m wide and 500-5000 m long. Ridges present a straight to crescent-shaped planform, with the outer limbs bent downflow in the latter case. Their cross-section is generally asymmetric, with a steeper lee face. They are very often associated to drumlins or megaflutings (Fig. 2.1), ribbed moraines occurring preferentially on topographic lows and drumlins on topographic highs. The progressive transition or superposition of these bedforms suggests a contemporaneous formation by similar subglacial processes (Aario, 1977a; Aario, 1977b). Like drumlins, their composition is highly variable but generally consists of material also found in their surroundings, commonly affected by glaciotectonic structures. For a clear review of their characteristics and different theories of formation, see Hättestrand (1997).

Formation

Drumlins and ribbed moraines are the most controversial bedforms. Although all theories agree that they formed subglacially at the end of the last glacial cycle, no agreement exists concerning their formation process, despite the abundant scientific literature dedicated to the question (Menzies, 1979; Menzies & Rose, 1987; Benn & Evans, 1998). Many of the theories fail to provide a general model of formation because they focalize on a particular bedform set.

Most of the recent theories recognize that flutings, drumlins and ribbed moraines are part of a continuum of subglacial bedforms and that their streamlined shape offers the least possible resistance to flow. Of course, bedform generation occurs at the interface between sand and water or air, leading to dunes or ripples. But it also occurs at other interfaces: between snow and air (Fig. 2.3 a-c) or between different masses of air, producing distinctive clouds (Fig.
As bedform-like morphologies occur at so many different interfaces, the question concerning the subglacial bedforms is to know at which interface they formed. Two main models exist: the subglacial deformation model and the meltwater hypothesis, favouring respectively the ice/bed and the meltwater/bed interface.

The subglacial deformation theory suggests that streamlined bedforms are the result of the deformation of a sediment layer located between the undeformed bed and the ice (Smalley & Unwin, 1968; Boulton, 1987). The theory is widely accepted but several variants exist. According to Boulton (1987), the coexistence of different lithologies (e.g., till and glaciofluvial gravels) in the subglacial deforming layer would lead to different deformation rates. Patches of less deformable sediments (e.g., gravels) would be the nuclei for the formation of drumlins. Ribbed moraines would represent the early stage of the drumlinization of transverse ridges, the preexisting transverse ridges being drumlins or megaflutings formed when the ice flow had another orientation. A more recent theory considers subglacial bedforms as instabilities at the interface between the deforming bed and the ice (Hindmarsh, 1998; Hindmarsh et al., 2003). Based on mathematical models taking into account their relative viscosities, computer-based simulations of drumlin formations have been investigated (e.g., Fowler, 2000; Schoof, 2002; Hindmarsh et al., 2003). The relevance of these highly mathematical studies is limited by the difficulty to evaluate the physical parameters of the subglacial environment. However, a similar approach for the formation of different types of aeolian dunes has led to good results with a simple algorithm (Nishimori & Tanaka, 2003). As dunes and drumlins share morphological characteristics, the simulation of drumlin formation could also be successful in the future.

According to the meltwater hypothesis (Fisher & Shaw, 1992; Shaw, 1998), streamlined bedforms were formed by catastrophic releases of meltwater forming subglacial sheet floods between the ice and the bed. One of the arguments is the striking resemblance between drumlins and erosional marks found at the base of turbidites (flute casts). Meltwater floods would create two types of drumlins (Shaw et al., 1989; Shaw, 2002): 1) erosional drumlins produced by the erosion of horseshoe vortices on previously deposited subglacial material (e.g., Shaw & Kvill, 1984; Shaw & Sharpe, 1987); 2) cavity-fill drumlins are formed by erosion of the overlying ice by meltwater and subsequent infilling of these cavities by sediments, leading to the creation of “inverted erosional marks” (Shaw, 1983; Sharpe, 1987). Ribbed moraines would be inverted erosional marks, like
cavity-fill drumlins formed by the infill of ripple-like cavities under the ice. This model presents an analogy with the formation of transverse ridges, which are giant ripple-like structures directly eroded into the bedrock (Shaw, 2002). The recent publication of bathymetric data from the previously glaciated Norwegian shelf showing giant ripple-like ridges (Ottesen et al., 2005) adds sense to the ribbed moraine formation by meltwater.

Other theories consider ribbed moraine as glaciotectonic products without genetic link to the associated drumlins, formed later. The debris-rich ice model (Shaw, 1979; Bouchard, 1989) proposes a formation by glaciotectonic shearing and stacking of debris-rich basal ice due to compressive ice flow. Englacial debris dragged along subglacial shear planes would then be released by basal melt-out. Transition to drumlins is interpreted as a transition from compressive to extending flow at the glacier bed. The fracturing model (Hättestrand, 1997; Hättestrand & Kleman, 1999; Kleman & Hättestrand, 1999) suggests that ribbed moraines result from the fracturation and “boudinage” of frozen till sheets due to extensional forces at the base of the ice, each ridge being part of the fractured till sheet. Extensional forces required by the model would result from the transition from a cold-based (ice fixed to the substratum) to a warm-based thermal regime (ice sliding over the substratum through a water film) during ice retreat.

2.2.2 Eskers

Eskers are elongated, sinuous ridges of glaciofluvial sands and gravels resulting from the infilling of subglacial, englacial or supraglacial ice-walled river channels, reaching lengths of hundreds of kilometres (Shreve, 1985). Esker path is controlled by the hydraulic gradient at the base of the glacier, which is itself controlled primarily by ice surface topography and, to a lesser extent, by the glacier base topography (Shreve, 1972). The effect of ice slope is 10 times more important than the conduit slope. Consequently, esker direction shows a close correspondence with the most recent direction of regional ice movement. Eskers sands and gravels are generally arranged in sequences recording fluctuations in discharge and sediment availability. Most of the eskers visible today formed in a subglacial ice-tunnel (R-channel) under inactive ice at the end of the last glacial cycle, without subsequent reworking by ice movement. The infill of the subglacial channel requires a significant decrease of meltwater flow and occurs therefore after the peak of meltwater discharge. Large eskers would form asynchronously near the ice margin, growing progressively upflow during ice retreat to form finally a single large landform. For more informations about eskers, see Banerjee & McDonald (1975) and Brennand (2000). For theoretical considerations about the englacial/subglacial meltwater system, see Roethlisberger (1972).
Chapter 3

Methods

3.1 Altimetric data

A digital elevation model (DEM) is a digital file that contains only spatial elevation data as a regular gridded pattern in raster format. Three different DEMs were used in the present study, varying in spatial extent (Fig. 3-1) and resolution (Table 3-1, Fig. 3-2).

SRTM (Shuttle Radar Topography Mission) data were acquired by radar interferometry from the space shuttle Endeavour at an altitude of 233 km during an 11-day mission in February 2000 (Rabus et al., 2003). This mission obtained elevation data on 80% of the Earth landmass (between 60°N and 54°S) with a spatial resolution of 1 arc-second (~ 30 m at the equator). The absolute and relative vertical accuracy are respectively ± 16 m (for 90% of the data) and ± 6 m on a local, 50-100 km scale (Rabus et al., 2003). The resulting DEM is a surface model: elevations refer to the top of the visible surface, including vegetation and buildings and not to the true ground elevation. Because of the limitations of the radar technique, water surfaces are irregular. A version of the entire dataset resampled to 3 arc-seconds, known as SRTM-3, is freely available (CIAT, 2005). The spatial resolution of 3 arc-seconds corresponds to cells of ~ 90 m at the equator. Towards the poles, the meridians get closer and the X-resolution increases accordingly. At the latitude of Switzerland, the resolution is 63 m in X and 90 in Y. The SRTM-3 horizontal resolution is sufficient to recognize the largest glacial landforms (e.g., drumlin fields) but insufficient to map them with precision (Smith et al., 2006). For large features, however, the SRTM-3 resolution and homogeneity are very useful.

ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) is another freely available DEM with a horizontal resolution 3 times better than SRTM (30 m instead of 90 m). Unexpectedly, some glacial features were visible on the SRTM but invisible on the ASTER, probably because of the greater inaccuracy in elevation of the latter.

The DHM25 (Digital Height Model 25) is the national DEM for Switzerland, consisting of the digitization of the
1:25'000 maps contour lines and their interpolation into a grid with a cellsize of 25 m. At regional scale, the official vertical accuracy is between 1.5 m in flat areas and 3 m in Alpine areas (Marti, 2004), but relief smaller than the original contour lines spacing (5 or 10 m) are omitted. DHM25 is much sharper than SRTM data but presents two types of artifacts related to the interpolation between contour lines: the first one gives a staircase appearance to the slopes, each step corresponding to one of the original contour lines, which is particularly inconvenient for the calculation of local slope or curvature; the second one consists in horizontal and vertical bands particularly visible in flat areas. Because the DHM25 is based on contour lines, most anthropogenic features are absent, unlike the Lidar or SRTM data which are based on aerial or spatial laser measurements.

Airborne LiDAR (Light Detection And Ranging) is a high precision remote sensing technology, able to cover large areas rapidly at a relatively low cost. It sends pulses of laser light towards the ground and detects the return times of back-scattered energy in order to determine the distance to the surface. LiDAR systems consist of three basic components: the laser scanner, the Inertial Measurement Unit (IMU), and a kinematic Global Position System (GPS). The last two components determine the aircraft’s position and orientation. Data are recorded in flight and are later post-processed to return a coordinate file (x,y,z) of the ground surface (Whitman, 2002). Additional data analysis and filtering allows production of a digital surface model (DSM) including vegetation and buildings and a digital terrain model (DTM) representing the ground elevation (Riedo et al., 2001). Finally, irregularly-spaced points are interpolated into a regularly-spaced grid to produce a DEM. In Switzerland, acquisition of LiDAR data started in the year 2000 for elevations below 2000 m. The altimetric precision (simple standard deviation) of the terrain model is approximately of 0.25 m and the original point density is higher than 1 point by m$^2$ in open areas. For the present study, we have had access to the LiDAR DTM for the cantons of Geneva and Vaud (cell size: 1m, resampled to 5 m for easier handling) and to the LiDAR DSM for part of Haute-Savoie (France) (cell size: 20 m). The Haute-Savoie DSM is affected by faint, 100 m-spaced NW-SE trending, linear artefacts, which correspond to the boundaries between the acquisition sweeps.

Successful landform identification depends heavily on the DEM scale (Smith et al., 2006). Any digital recording of a continuous signal requires sampling this signal at a regular interval, which corresponds to the DEM cell size in the case of the earth surface. The smallest recordable wavelength is the Nyquist wavelength, which corresponds to the length of two sample intervals. Consequently, only morphological features larger than a square of 2 x 2 cells are reliably recognizable. For the entire Swiss territory, the DHM25 25m-resolution does not permit to recognize features smaller than 50 m and certainly misses some glacial landforms. For Geneva and Vaud, the LiDAR DEM 1m-resolution permits to recognize all features larger than 2 meters, i.e., all the glacial landforms.

<table>
<thead>
<tr>
<th>DEM</th>
<th>Resolution</th>
<th>Acquisition date</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shuttle Radar Topography Mission (SRTM)</td>
<td>X: 63 m, Y: 90 m</td>
<td>February 2000</td>
</tr>
<tr>
<td>Digital Height Model 25 (DHM25)</td>
<td>25 m</td>
<td>digitization of swiss 1:25'000</td>
</tr>
<tr>
<td>Light Detection And Ranging (LiDAR) Geneva &amp; Vaud</td>
<td>1 m (resampled to 5 m)</td>
<td>GE: May 2000; VD: October 2002</td>
</tr>
<tr>
<td>Light Detection And Ranging (LiDAR) Haute-Savoie</td>
<td>20 m</td>
<td>unknown (~ 2000)</td>
</tr>
</tbody>
</table>

Table 3-1 Resolution and acquisition date of the DEMs used in this study.
3.2 Coordinate system

All the maps presented in this work use the Swiss coordinates system (or Swiss Grid), an oblique Mercator projection based on the 1841 Bessel ellipsoid. CH1903, the geodetic datum, is measured in meters and has its fundamental point (600'000 m E / 200'000 m N) located in Bern (7°26'22.5"E / 46°57'08.66" N) and its 0 / 0 point near Bordeaux (France). This datum leads to the following advantages on the Swiss territory: 1) coordinates are positive; 2) X coordinates are always larger than Y coordinates; 3) coordinates contain 6 digits when measured in meters. In this work, coordinates are indicated in kilometers (xxx.x / yyy.y) for brevity.

3.3 Topographic attributes and representation

DEM's are uniformly sampled grids of elevation values representing a continuous surface. To obtain a geomorphological map from it, landforms must be delimited according to the break of slope at their perimeter. Optionally, a line can indicate their crest. For these tasks, many attributes can be derived from the DEM data, enhancing different aspects of the landscape (Wood, 1996; Shary, 2002). Below are presented the different topographic attributes used in this work (Figs. 3-3 and 3-4) and their utility for landform mapping. Most of them were generated in ArcGIS 9 with the Spatial Analyst extension. For practical details concerning their calculation, see appendix A.3.

Contour mapping is the traditional representation of a 3D topographic surface on a 2D medium and can be generated easily from a DEM. It implies drawing contour lines linking the points of equal elevation for a given elevation interval. By choosing a contour interval much smaller than the landform height, one can recognize easily breaks of slope by the transition from closely spaced contours on the landform sides to more widely spaced contours around the landform. This representation method is accurate but presents a major shortcoming: one has to read the values of the contour lines to know the slope direction and to avoid confusion between concave and convex morphologies. Besides, a contour map does not provide an intuitive view of the landscape and requires experience to obtain a correct mental representation of the topography.

Elevation display is the most direct DEM representation, in which a color scale corresponds to the DEM elevation range. For the detection of glacial landforms, this method of representation is applicable to areas of low relief (i.e., a low range of elevations), where the present topography reflects mainly the Quaternary process (e.g., Curtis, 1996 for the lowlands of Illinois). Mountainous areas like Switzerland present an elevation range much greater than the glacial landforms and the latter are drown within the regional topography. For instance, the case for instance between Geneva (372 m) and the foot of the Jura foothills (ca. 700 m): whereas the elevation range is around 330 m, superimposed moraines have a maximum amplitude of 20 m and are thus undetectable for the human eye on a mere elevation display. When zooming on a small portion of the DEM, however, the elevation range decreases and the elevation display can give a good representation of the glacial features.

Low-cut elevation is obtained through low-cut spatial frequency filtering (also called high-pass filtering), which suppresses the regional topographical variations (large spatial wavelengths, low spatial frequency) and preserves only the local variations (short spatial wavelengths, high spatial frequency), including the glacial landforms in our case. The drawback of this method is that low-cut filtered discontinuities (e.g., terraces borders, riverbanks) in the landscape may appear as ridges hardly distinguishable from original ridges (e.g., eskers, drumlins, moraines).

Hillshading (sometimes also called reflectance) simulates an illumination of the ground. It is the most widely used DEM representation method because it provides a very realistic and detailed view of the terrain. Vertical exaggeration of the topography can significantly enhance the visibility of subtle features. However, the method suffers from its directional illumination, which overestimates illumination-perpendicular structures and underestimates parallel ones. In order to overcome this azimuth biasing, it is important to use at least two different hillshaded maps for interpretation, preferentially perpendicular and parallel to the main landforms orientation (Smith, 2003). If not specified, the hillshaded views presented in this study have an illumination azimuth of 315° (NW), an illumination elevation of 45° and a vertical exaggeration of 2.

Slope expresses the local slope of the DEM. It is equivalent to a hillshading with the light source located at the zenith. This particular illumination does not produce azimuth biasing. The disadvantage of slope is the possible confusion between convex and concaves morphologies.

Profile curvature (also called slope curvature or concavity) is the curvature of the earth surface, measured in the vertical plane parallel to the slope direction. It indicates the rate of change of slope in the slope direction, corresponding to the DEM second derivative. The advantages of the profile curvature over hillshading are the absence of azimuth biasing and the highlighting of the landforms crests and breaks-of-slope, which are both useful for landform mapping. In the case of a drumlin cross-section, the slope changes from flat at the base, moderate on the sides and flat again at the top. Consequently, curvature is strongly concave (positive) at the borders and strongly convex (negative) at the top. For an intuitive representation, curvature should be displayed with a white to black color scale corresponding respectively to convex and concave features, because concave features are usually less exposed to sunlight than convex features in a natural environment.

Positive and negative openness are two recently published topographic attributes expressing respectively “the degree of dominance or enclosure of a location on
Fig. 3-3 Several attributes of the DHM25 DEM for the Zurich highland drumlin field area. a) raw elevation; b) low-cut elevation; c) hillshading (vertical exaggeration: 2x, light source: NW, 45°); d) hillshading (vertical exaggeration: 2x, light source: NE, 45°); e) positive openness; f) slope; g) profile curvature; h) low-cut elevation blended with slope.
Fig. 3-4 Several attributes of the 5m_LiDAR DEM for the Bois-de-Chêne area. a) raw elevation; b) low-cut elevation; c) hillshading (vertical exaggeration: 2x, light source: NW, 45°); d) hillshading (vertical exaggeration: 2x, light source: NE, 45°); e) positive openness; f) slope; g) profile curvature; h) low-cut elevation blended with slope.
an irregular surface” (Yokoyama & Shirasawa, 2001; Yokoyama et al., 2002). Positive openness is an angular measure of the area of sky visible from the each cell location, within a given perimeter (Fig. 3-5a). It provides a dramatic representation of the landscape where largest values corresponding to crest or peaks and lowest values to the bottom of steep-sided valleys. For the present study, positive openness was generated for a diameter of 1000 m and proved to be very efficient to highlight the crest of glacial landforms such as eskers, drumlins and moraines.

For glacial landform mapping, we used a combination of all the topographic attributes described above (except contour maps), taking advantage of the ability of ArcGIS to switch rapidly between different layers and to blend them. Our experience confirms the work of Smith (2003), who showed that no single DEM attribute is able to provide an optimal viewing for glacial landform mapping. Their advantages and disadvantages are summarized in Table 3-2.

For a first identification of the glacial landforms over a large area, hillshading or openness are the best attributes because they provide the most intuitive perception of the relief. Openness is also adapted to the mapping of landform crests (particularly in the case of streamlined features like drumlins).

In order to map the perimeter of large landforms, slope or profile curvature are ideal because they highlight the breaks of slope surrounding the landform and do not suffer from azimuth biasing. However, profile curvature is not adapted for low, subtle landforms: on the DHM25, the original contour lines used for the DEM creation produce artificial steps on the slopes. Cut transversally, these artifacts create an alternance of high concave and convex values and these may mask the natural convexity/concavity of the landscape. Concerning the higher-resolution LiDAR DEM, profile curvature is sometimes too “noisy”: the curvature at the border of roads or buildings exceeding much the curvature of the smooth glacial features. When the landforms ares not limited by a clear break of slope, profile curvature or slope are not adapted and mapping is based on raw elevation or low-cut elevation.

Topographic attributes can be represented with a greyscale or a colour scale. The greyscale is more objective than the colour scale because the transition from black to white is continuous and takes advantage of the human eye sensitivity to lightness variation. On the contrary, the transition between the different colors of a rainbow-like colour scale appears irregular because of the nonlinear perception of wavelength change, providing the sensation

<table>
<thead>
<tr>
<th>Topographic attribute</th>
<th>Advantages</th>
<th>Disadvantages</th>
</tr>
</thead>
<tbody>
<tr>
<td>Elevation display</td>
<td>Visualisation of the regional trend. Pure elevation information.</td>
<td>Poor rendering of local relief.</td>
</tr>
<tr>
<td>Low-cut elevation display</td>
<td>Removes the elevation regional variations and extracts the local relief.</td>
<td>Depends on the moving window size. Possible confusion between ridges and topographical steps.</td>
</tr>
<tr>
<td>Hillshading</td>
<td>Provides a very intuitive and detailed view of the landscape.</td>
<td>Highlighted features depend on the light orientation.</td>
</tr>
<tr>
<td>Slope</td>
<td>Provides and intuitive and detailed view of the relief without azimuth biasing.</td>
<td>Potential confusion between convex and concave landforms.</td>
</tr>
<tr>
<td>Positive Openness</td>
<td>Highlights the crest of the ridges. Intuitive view of the landscape.</td>
<td>Ridges appear narrower</td>
</tr>
<tr>
<td>Profile curvature</td>
<td>Highlights the breaks-of-slope limiting the landforms.</td>
<td>Dependent on the moving window size.</td>
</tr>
<tr>
<td>Contour lines</td>
<td>Renders both regional trends and local details with accuracy.</td>
<td>Potential confusion between convex and concave landforms.</td>
</tr>
</tbody>
</table>

Table 3-2 Comparison of the DEM attributes advantages and disadvantages.
Chapter 3 - Methods

that some parts of the colour scale change more rapidly than others (Gregory, 1966). However, the association of greyscale and colour scale is extremely useful to represent two different attributes simultaneously, by transparency. In the present work, the elevation or low-cut elevation was represented with a colour scale and blended by transparency with a greyscale representation of a second attribute highlighting the breaks of slope such as hillshaded view, slope or profile curvature (Figs. 3-3 h and 3-4 h). The association of low-cut elevation and slope is particularly useful as the slope attribute offers an intuitive, detailed, non-azimuth-biased perception of the relief, while the low-cut elevation avoids the possible confusion between concave and convex features.

3.4 Landform mapping

The classification of a continuous surface (topography) into discrete entities (landforms) is to a certain extent subjective, but this operation is necessary to obtain morphological values such as elongation ratio or landform height. These values permit the comparison of landforms within a given field or between different areas (e.g., Alpine vs North American drumlins).

In the investigation of former glaciations, landform mapping has always been an essential method mainly based on observations in the field, on topographic maps or on aerial photographs. The digital mapping of landforms based on a DEM presents several advantages over these techniques: easy navigation between the regional and local scales, digital processing of the DEM to extract useful surface attributes and automatic calculation of statistics concerning the landform morphometry. Besides, the resulting shapefiles are geo-referenced and ready to be integrated in any GIS.

Ideally, drumlins resemble half-buried eggs and are delimited by the line down to which the egg is buried (the break of slope). This line often corresponds to a contour line, but this is not the case when drumlins rest on a slope rather than on a horizontal surface. In areas where the drumlins are particularly well developed, automated methods can be used to recognize drumlin shapes on a DEM, leading to a more objective mapping (Munro-Stasiuk, 2005). The use of such methods is limited in Switzerland for several reasons: irregular drumlin shape, association of drumlins with ribbed moraines, progressive fade-out of the drumlin lee-side without clear break of slope, contiguity of two drumlins, smoothing of the drumlins by overlying lacustrine deposits and finally partial erosion by post-glacial streams.

Two types of drumlin maps exist: those representing the crest of drumlins and other glacial lineaments with a single straight line (e.g., Clark & Meehan, 2001) and those indicating the contour, crestline and highest point of each drumlin (e.g., Riley, 1987). We adopted the first method for rock drumlins, due to their poorly defined contour, and the second for standard drumlins.

The contour and/or crest of glacial landforms were digitized manually, through visual interpretation, in shapefiles, a shapefile being a set of files containing georeferenced vector features (either points, polylines, or polygons) and associated tabular data. These are listed in Table 3-3.

<table>
<thead>
<tr>
<th>Landform</th>
<th>Shapefile name</th>
<th>Shapefile type</th>
<th>Original DEM</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rock drumlins</td>
<td>Molasse lineaments dhm25.shp</td>
<td>polyline</td>
<td>DHM25</td>
<td></td>
</tr>
<tr>
<td>Molasse morpho facies</td>
<td>molasse_MorphFacies.shp</td>
<td>polygon</td>
<td>DHM25</td>
<td></td>
</tr>
<tr>
<td>Drumlin fields</td>
<td>Drumlins_fields.shp</td>
<td>polygon</td>
<td>SRTM, DHM25, LiDAR</td>
<td>Name stored in field “name”</td>
</tr>
<tr>
<td>Drumlins</td>
<td>drumlins.shp</td>
<td>polygon</td>
<td>DHM25, LiDAR</td>
<td>For table fields, see Fig. 5-2</td>
</tr>
<tr>
<td>Dombes area ridges</td>
<td>Dombes_ridges.shp</td>
<td>polyline</td>
<td>SRTM</td>
<td>For table fields, see Table 6-1</td>
</tr>
<tr>
<td>Ribbed moraines</td>
<td>RibbedMoraines.shp</td>
<td>polygon</td>
<td>DHM25, LiDAR</td>
<td></td>
</tr>
<tr>
<td>Jura Till patches</td>
<td>joux_till_patches.shp</td>
<td>polygon</td>
<td>LiDAR</td>
<td></td>
</tr>
<tr>
<td>Joux valley esker</td>
<td>joux_esker.shp</td>
<td>polygon</td>
<td>LiDAR</td>
<td></td>
</tr>
<tr>
<td>Bois-de-Chêne eskers</td>
<td>BDC_esker.shp</td>
<td>polyline</td>
<td>LiDAR</td>
<td></td>
</tr>
<tr>
<td>Moraines</td>
<td>moraines.shp</td>
<td>polyline</td>
<td>DHM25, LiDAR</td>
<td>Id=0: minor moraine</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Id=1: major moraine</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Id=2: Montosset moraine</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Id=3: Jura ice cap moraine</td>
</tr>
<tr>
<td>Paleo rivers (Bière area)</td>
<td>Biere_paleo_rivers.shp</td>
<td>polygon</td>
<td>LiDAR</td>
<td>Id=0: Paudex gravels</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Id=1: Paleo-Serine</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Id=2: Paudex</td>
</tr>
</tbody>
</table>

Table 3-3 Shapefiles concerning glacial landforms, available on the CD-rom (Appendix F)
Quaternary glaciations have led to an important overdeepening in all perialpine valleys, reaching depths below present sea level in many places (Finckh et al., 1984). Most of the Swiss lakes are located in these valleys. In the north-alpine valleys, the overdeepening is attributed to glaciations, even though shallower pre-Quaternary valleys must have existed and guided glacier and meltwater flow. The south-alpine valleys present deeper incisions, because of the catastrophic drop in the Mediterranean sea level during the Messinian crisis. A new map of the bedrock elevation in Western Switzerland, compiled from published and new material is presented (Fig. 4-1).

4.1 Methodology

The bedrock elevation of western Switzerland has been investigated by several authors. These separated data sources were compiled into a single geo-referenced digital grid, providing a more global and accurate view of the bedrock morphology for the investigated area. Besides, the digital format permits the derivation of surface attributes (e.g., slope, hillshading) and the integration with other georeferenced data sources.

The main data source is the pre-Quaternary bedrock depth of Western Switzerland of Pugin (1988), which completed that proposed by Wildi (1984) in eastern Switzerland. Pugin’s map reports bedrock elevation up to a maximum height of 750 m, based on boreholes, seismic reflection and gravimetry. It was extended and refined with four other datasets, including new seismic data from the present work (Table 4-1, Fig. 4-2).

Datasets 1 and 3 were digitized using the method described in appendix A.1. For dataset 4, only the original seismic information was digitized (i.e., value of isochrones at intersection with seismic lines), the easternmost part of the lake (east of x coordinate 545.0) being omitted, where seismic energy did not reach the deep bedrock, because of the thick sediment accumulation of the Rhone delta. For seismic datasets 4 and 5, seismic reflection times were converted to depth using a velocity of 1700 m/s and to elevation based on a lake level of 372 m. Finally, the grid was created by interpolation of the five datasets in ArcGIS, using a cell size of 50 m.

By subtracting the bedrock elevation from the present topography (DHM25), an isopach map of the Quaternary deposits was obtained. The area covered by the map contains several major lakes (Geneva, Neuchatel, Murtensee, Bielersee and Thunersee) and the DEM had to be corrected to indicate the elevation of lake bottom rather than lakes surface. The data processing is illustrated in figure 4-3.

4.2 Bedrock morphology

The bedrock elevation map offers a clear representation of the glacial valley network (Fig. 4-1), which has a different morphology between the Alpine range and the Molasse Plateau.

<table>
<thead>
<tr>
<th>#</th>
<th>Dataset</th>
<th>Geographical Zone</th>
<th>Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Pugin (1988)</td>
<td>Western Switzerland</td>
<td>gravity, boreholes, seismic</td>
</tr>
<tr>
<td>4</td>
<td>Dupuy (2005)</td>
<td>Grand Lac (eastern Lake Geneva)</td>
<td>Seismic reflection</td>
</tr>
<tr>
<td>5</td>
<td>Present work</td>
<td>Petit Lac (western Lake Geneva)</td>
<td>Seismic reflection</td>
</tr>
</tbody>
</table>

Table 4-1 Data used for the bedrock map.
Fig. 4-1 Map of bedrock elevation in Western Switzerland. Elevation-dependent color scale blended with slope.
Fig. 4-2 Presentation of the data sources used to generate the bedrock elevation surface of Western Switzerland.
Chapter 4 - Bedrock morphology in Western Switzerland

Digitization of the lakes contour lines (ArcScan)

Scanned 1:25000 Swiss map for the Lakes

Shapefile of the lake bottom contour lines

Conversion to Raster (Spatial Analyst / Topo to raster)
  Boundary: Lake borders
  Cellsize: 50 m

Grids of lake bottoms

Mosaicing of the lakes on top of the DEM
  Cellsize: 25 m

Lake water-free DEM

Bedrock grid

Subtraction (DEM – Bedrock)

Quaternary isopach map

Legend:
  Data
  Process

Fig. 4-3 Flow chart for the methodology used to derive the Quaternary isopach map.

Fig. 4-4 Map of bedrock elevation in the Geneva area. Elevation-dependent color scale blended with slope. Black line indicates present Lake Geneva shoreline. Note the valley anastomosing pattern and undulating longitudinal profile. The only exception is the Allondon valley: its longitudinal profile presents a regularly dipping slope (~0.7%) typical of a subaerial stream.
Chapter 4 - Bedrock morphology in Western Switzerland

In the Alpine range, alpine valleys form a dendritic pattern where small valleys merge into larger ones, the largest ones being the Rhone and Aar valleys. Their orientation is NW-SE, perpendicular to the tectonic trend. They are steep-sided, V-shaped and deeply cut the bedrock, down to 400 m below sea level at the eastern end of Lake Geneva. These characteristics indicate that meltwater drainage occurred as a single channel located at the bottom of the main valley.

On the plateau, valleys are oriented SW-NE, parallel to the Alpine front. They are shallower, wider and anastomosed in places, with an undulating or concave longitudinal profile. These characteristics are particularly well identified around Geneva (Fig. 4-4) thanks to the high density of boreholes and seismic lines (Signer, 1996). Valleys in the Geneva area show a typical width of 1 km but range from 200 m in constricted segments to more than 2 km in the Petit-Lac (western Lake Geneva). The latter extends southwestwards into the Aire river plain and constitutes the deepest valley in the Geneva area. These characteristics indicate that meltwater drainage occurred as a single channel located at the bottom of the main valley.

The change in morphology between the Alpine range and the plateau reflects the transition from alpine to Piedmont glaciers. The plateau valleys have a flat bottom, an anastomosed pattern and undulating longitudinal profiles with up-slope segments. These characteristics are typical of networks cut into bedrock or sediments by subglacial meltwater are known as “tunnel valleys” (O’Cofaigh, 1996), where meltwater flows at hydrostatic pressure. Therefore, the subglacial meltwater drainage pathways depend primarily on the elevation of the ice surface and, to a much lesser extent, on the bed topography. This explains that their longitudinal profile typically crosses topographic barriers.

When the ice covered completely the Molasse plateau between the Prealps and the Jura, the Petit Lac valley probably constituted the main tunnel valley, draining a large amount of meltwater. When the ice volume was smaller, the Rhone ice tongue used this valley and widened it, but the bottom of the valley was still occupied by the meltwater drainage.

The network of tunnel valleys was formed by meltwater activity during the numerous Pleistocene glaciations. Was this network active during the WGM? In the Geneva area, tunnel valleys are filled by a gravel unit (sandur) known as Alluvion Ancienne, deposited before the WGM (see section 9.3 for a presentation of the Geneva area Quaternary stratigraphy). During the WGM, the Alluvion Ancienne was completely eroded by meltwater in the Aire valley (the prolongation of the Petit Lac) but was partly or completely preserved in other smaller tunnel valleys (Figs. 4-4, 4-5 and 4-6). This shows that the subglacial meltwater circulation during the WGM was less important than during previous glacial periods, during which the whole tunnel valley network was used. This smaller meltwater activity during the WGM is also supported by

Fig. 4-5 Hillshaded view of the Geneva and Haute-Savoie Lidar DEMs showing the prolongation of the Petit-Lac tunnel valley through the Geneva area.

Fig. 4-6 Upper surface of the Alluvion Ancienne in the Geneva area (Jaquet & Kaufmann, 2002). Note the absence of the Alluvion Ancienne in the southern onshore prolongation of the Petit Lac valley (Aire Valley). Other large areas without Alluvion Ancienne correspond to bedrock hills or to areas without sufficient borehole information.

Fig. 4-7 Interpreted photo-terrestrial map showing the prolongation of the Petit-Lac tunnel valley through the Geneva area.
the preservation of older sediments (Riss) in several parts of Switzerland.

4.3 Quaternary isopach map

Figure 4-7 presents the Quaternary isopach map for Western Switzerland. Over the 4000 km$^2$ investigated, Quaternary sediments represent a volume of 300 km$^3$, equivalent to an average thickness of 75 m. The Quaternary isopach grid presents anomalous negative values where the bedrock grid is higher than the DEM. This error comes from the lower level of precision of the bedrock map relative to the DHM25. As expected, these deposits were accumulated mainly in the tunnel valleys. The latter were occupied by lakes as glaciers retreated. Since then, these lakes have been filled partly or completely by sediments, mainly by glacio-lacustrine clayey silts (Pugin, 1988; Hinderer, 2001). Similar deposits fill the tunnel valleys of northwestern European lowlands (Huuse & Lykke-Andersen, 2000; Huuse et al., 2003) and in the North Sea (Praeg, 2003).

The greatest Quaternary accumulation corresponds to the Rhone delta, with a maximum thickness of 850 m. The important overdeepening of the bedrock valley combined with the high sedimentation rate of the delta explains this accumulation. Apart from the tunnel valley infills, the only other important Quaternary accumulation corresponds to the “graviers de la Côte” and Bière area, north of Rolle.
Chapter 5

Rock Drumlins in Western Switzerland

5.1 Introduction

During the WGM, the Molasse foreland between Lausanne and Solothurn was covered by the Rhone glacier's northeastern arm. Guided by the Alpine front and the Jura range, the general ice flow direction changed down-ice from north to northeast. Today, the Molasse gentle relief on the Swiss Plateau is characterized by numerous elongated hills (Fig. 5-1).

Below a thin and discontinous Quaternary till cover, hills are composed of Chattian-Aquitanian Lower Freshwater Molasse (USM, sandstones and marls deposited in an alluvial environment) and Burdigalian Upper Marine Molasse (OMM, marine sandstones). Each unit is subdivided in formations with variable proportions of clay, sand and limestone. Because the Molasse results from the erosion of the Alps, grain size decreases away from the Alpine front.

The Molasse structure consists mainly of broad, low-amplitude anticlines and synclines oriented SW-NE (Fig. 5-2). Between Lausanne and Yverdon, the Molasse presents an isoclinal structure dipping gently towards the SE, except in the eastern Jorat upland (Fig. 5-1), where it becomes subhorizontal. The Alpine front is bordered by a zone of thrusted Molasse known as subalpine Molasse.

5.2 Previous studies

Previous studies on the elongated Molasse hills have been conducted principally in the canton of Vaud, between Lausanne and Yverdon. There, the formation of the SSW-NNE elongated Molasse hills are clearly related to the differential erosion of the thick competent sandstone and softer marls beds. Little information exists about the exact ice flow direction, because glacial striations were mapped only in the Lausanne area (Choffat & Aubert, 1983). Because the ice flow was sub-parallel to the tectonic trends, the relative influence of these two elements in the hill orientation has been a matter of debate since the beginning of the last century.

Biéler (1901) considered these hills as a typical drumlin landscape resulting from both Molasse erosion and till deposition. For him, all the elongated crests were drumlins, including the prominent conglomerate beds of the Mont Pélerin. He noted their longitudinal asymmetry (steeper stoss side) and their preferential formation where ice flow was parallel to the tectonic trend. On the contrary, Bersier (1942) proposed that drumlins were totally absent in the area, where the relief consisted only of cuestas formed by meteoric and fluvial erosion, without glacial action. Bersier's main argument was the absence of streamlined landforms perpendicular to the tectonic trend. To explain the relatively short extension and poor alignment of the individual cuestas, he invoked lateral variations in lithology/competence within the Molasse beds and small transverse strike-slip faults. Aubert (1981) proposed an intermediate interpretation: the numerous elongated Molasse hills were shaped by ice but were not drumlins because their orientation reflected the tectonic trend, and not the ice flow.

Little attention has been paid to the elongated Molasse hills located further east, in the Seeland and Mittelland. Van der Meer (1982) mapped them in the Fribourg area and concluded that their elongation indicated the ice flow, independently of the tectonic trend. Although many drumlins have a tail of glacial material, drumlins formed entirely of glacial material seem absent in this area (van der Meer, 1982).

Similar streamlined bedrock hills have been described in many formerly glaciated areas and called rock drumlins. Rock drumlins result from the abrasion of soft beds, preferentially limestones and sandstones and differ from the roches moutonnées, which result also from down-ice plucking of hard and jointed rocks (Knight, 2003). Based on examples from northern Scotland and Greenland, Gordon (1981) concluded that rock drumlins reflect bedrock structure regardless of ice flow direction. Similarly, Linton (1963) and Knight (2003) noted that they are only developed when ice flow is parallel to bedrock strike. However, rock drumlins do
Fig. 5-1 NW-hillshaded views of the DHM25 and Lidar DEMs showing the streamlined Molasse surface between Lausanne and Solothurn; a) global view (DHM25), b) alignment of flow-parallel rock drumlins along a flow-transverse Molasse sandstone bed (DHM25), c) area between Lausanne and Yverdon (LiDAR), d) Mt Pélerin area (LiDAR).
exist perpendicular to bedrock strike in certain areas (e.g., Dionne, 1987; Grosswald et al., 1992), allowing in this case a clear distinction between drumlins and bedrock structure. In any case, their occurrence and morphology are strongly related to bedrock structure.

### 5.3 Method

In order to evaluate the impact of the Molasse lithology and structure on the hill morphology, orientation and distribution, the morphological facies (morphofacies, Fig. 5-3) and linear topographic features (lineaments, Fig. 5-2) were mapped from the DHM25 DEM and compared with the Molasse lithology and structure, obtained from the existing geological maps (Fig. 5-2).

### 5.4 Results

#### 5.4.1 Mont Pélerin area

East of Lausanne, the Mt Pélerin area (Fig. 5-1d) consists of a Chattian coarse-grained alluvial fan deposits formed along the active Alpine front. Landscape morphology is characterized by prominent and elongated ridges (morphofacies MP in Fig. 5-3). Around the Mt Pélerin, the ridges are curved and follow approximately the contour lines. Theses ridges correspond to the intersection between the steep topography and the thick, sub-horizontal conglomerate layers. East of Mt Pélerin, the topography is conformable with the conglomerate layers and ridges are absent. Southeast of Mt Pélerin, between Vevey and Montreux, the layers strike (NE) is perpendicular to the
Chapter 5 - Rock drumlins in Western Switzerland

ice flow direction (NW) and ridges are poorly developed. These observations show that ridges of the Mt Pélerin area result from the carving of softer intervals between the conglomerate beds and that this differential erosion is much more important where the ice flow is sub-parallel to the structural trend.

5.4.2 Lausanne - Yverdon

The USM between Lausanne and Yverdon (Fig. 5-1c) consists of an alternation of sandstones and marls (Jordi, 1995). Elongated hills are irregularly spaced (morphofacies M2) and have a variable lithology. Their cross-section is asymmetrical with a steeper eastern side and a gentler western side, reflecting the Molasse isoclinal structure. Their longitudinal section presents generally a whaleback shape, the typical drumlin longitudinal asymmetry (steep stoss side, gentle lee side) being uncommon. Hills sometimes present a bended crest inherited from the Molasse structure. In some places, discordant orientations reflect stratigraphic discordances within the Molasse. Such a case is encountered south of Dizy (VD) (527.0/163.7, Fig. 5-4a). The discordance in orientation has an azimuth of 20°, parallel to the large bed on its western side. The eastern beds have a strike of 10° and pinch-out against the angular unconformity. Despite their difference in orientation, Molasse beds were smoothed on both sides by ice flow. This indicates that the rock-drumlin orientation between Lausanne and Yverdon do not reflect the ice flow but the orientation of the Molasse beds. In the northeastern part, southeast of Yverdon, elongated hills are absent or very low (morphofacies M1), reflecting a more homogenous

Fig. 5-3 Morphological facies map for the Swiss Plateau between Lausanne and Solothurn. In the legend, facies starting with M refer to the Molasse. MS: smooth; M1/M2/M3: low/medium/strong lineation; MP: strong lineation in the Mt Pélerin area; MU1/MU2: slightly/highly uneven, without lineations.
stratigraphy. The LiDAR data does not show any glacial lineation transverse or oblique to the stratification.

Several authors have reported tails of glacial material down-ice of Molasse beds. Unfortunately, the LiDAR DEM does not permit to differentiate these tails from the Molasse body. Such a situation occurs south of Dizy (527.2/164.5): at the downflow extremity of a prominent Molasse bed, a gravel pit reveals the presence of glacial material preserved from erosion by the up-ice Molasse bed (Fig. 5-4). On the high-resolution Lidar DEM, the boundary between the Molasse and the till is invisible. Other till tails probably exist where ice flow parallels the bedrock strike, for instance in the Biolay-Orjulaz area (535.0 / 165.0) where hills are particularly long and present a pronounced longitudinal asymmetry.

Besides the Molasse drumlins, several authors have reported some drumlins formed of glacial material. As Molasse hills are generally covered by a till layer, the recognition of drumlins formed entirely of glacial material is impossible without boreholes or excavations. On the existing geological maps, the distinction between till drumlins and molasse hills is generally based on morphological rather than lithological criteria: small forms being generally reported as drumlins and longer forms as Molasse hills (“buttes molassiques”) or moraines (“vallums morainiques”). According to Biéler (1901), till drumlins may be constituted entirely of till or contain a Molasse core. According to Bersier (1942), till drumlins are rare and always associated to either a bedrock core or a bedrock protection upflow.

As for drumlins formed entirely of glacial or glaciofluvial material, the only published drumlin section is that of “Le Paradis” (534.2/163.5), east of Cossonay (Burri et al., 1968). It results from the erosion of the stratified proglacial gravels of Biolay-Orjulaz (dated 34’600 years BP, Weidmann, 1974) and is draped by a thin till layer.

![Fig. 5-4 Outcrops in the gravel pit of Dizy in January 2006, at the downflow extremity of a rock drumlin. A sharp erosion surface separates a lower glacio-fluvial unit (U1) consisting of poorly sorted, stratified sands and gravels from an upper clayey till unit (U2). a,b) Location map on the LiDAR DEM. The excavation front has moved towards the south since the LiDAR acquisition. c) Schematic geometry. d) Longitudinal outcrop; note the undulating erosion level dipping towards the south. e) Transverse outcrop. f) Zoom of the contact between U1 and U2.](image-url)
Gravel exploitation has altered the hill morphology and the DEM does not provide new relevant observations. Biéler (1901) reported another gravel drumlin near Montcherrand (529.7/176.7), but the LiDAR sharpness reveals instead an irregularly shaped interfluve.

### 5.4.3 Yverdon - Solothurn

Southeast of Yverdon, the transition from USM to OMM corresponds to a positive step in the topography and to a marked change in morphofacies (M1 / M3, Figs. 5-1c and 5-2). Elongated hills become more numerous, less spaced, less prominent and less elongated. Following the eastern shore of Lake Neuchâtel towards the NE, the correlation between morphofacies and lithology is confirmed by the transition from OMM (M3) to USM (M1). This difference in morphofacies results apparently from the absence of marl intervals in the OMM, which consists mainly of marine sandstones. Where the OMM structure is conformable with the topography (e.g., in the Jorat upland, 545.0 / 160.0), the morphofacies is irregular (MU2) and lacks elongated hills.

In the Fribourg area (175.0 / 570.0) elongated hills diverge and are oriented perpendicularly to the N-S tectonic trend (Figs. 5-1, 5-2 and 5-3). Although the ice flow was perpendicular to the OMM beds, elongated hills formed in the ice flow direction. These hills are smaller and less abundant than where ice flow and structures are parallel, but their presence indicates that the ice flow had more influence on the topography than the Molasse structure. Such a situation is even more straightforward at the centre of the plateau, where the ice flow was more powerful (Fig. 5-1d, coordinates 584.0/201.4): well-developed elongated hills with a SW-NE orientation are aligned in echelon along a prominent Molasse bed oriented SSW-NNE.

### 5.5 Conclusions and perspectives

The elongated hills between Lausanne and Solothurn are rock drumlins made of Molasse conglomerates, sandstones and marls streamlined by glaciations. In some cases, a tail of glacial material is deposited or preserved behind the Molasse hill but the lithological change is not visible on the DEM. Rock drumlins acquired their present shape through the successive Quaternary glaciations. The shape of the studied rock drumlins is a function of lithology – development is enhanced by an alternation of softer and harder lithologies – and angle between ice flow and stratification. Rock drumlins are best developed where ice flow is parallel or sub-parallel to the stratification, but their direction does not indicate the exact ice flow, as already noted by Aubert (1981). In the Fribourg area, where ice flow was perpendicular or clearly oblique to the Molasse orientation, rock drumlins also formed. In the latter case, the orientation of drumlins does reflect the ice flow.

In the future, access to the high-resolution LiDAR dataset for the entire area covered by the Rhone glacier northeastern ice tongue (i.e., cantons of Vaud, Fribourg, Neuchâtel, Bern and Solothurn) would permit to refine the geomorphological mapping. Compilation of subsurface information from boreholes would make possible the identification of till drumlins, and highlight the relationship between Molasse composition and morphofacies.
Chapter 6
Drumlins in Switzerland

6.1 Introduction

The study of drumlins in the Swiss northern Alpine foreland began with Früh (1896). Based on field observations and topographic maps, this author provided a map indicating the location, extent and general landform orientation of drumlin fields (Fig. 6-1). This mapping is restricted to eastern Switzerland and does not show the location or shape of individual drumlins. Despite the numerous studies on glaciations in Switzerland since then, little attention has been paid on the morphology and distribution of subglacial landforms at the national scale since then.

The Swiss Geological Atlas 1:25000 maps report the contours of most of the drumlins but this mapping is limited by the accuracy of the pre-existing topographical

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Fig. 6-1 The only published map of drumlins in the eastern Swiss Alpine foreland (Früh, 1896).
Chapter 6 - Drumlins in Switzerland

maps and depends on the drawing style of the author, because observations in the field hardly allow to correctly evaluate the shape of glacial features (Fig. 6-2). These maps suffer from missing landforms (too small or hidden by vegetation), wrong landform delimitation (especially for drumlins) and wrong landform attribution (e.g., confusion between drumlins, moraines and eskers). Generally, drumlins are drawn more rounded than they really are: some narrow elongated drumlins visible on the DEMs appear as broad ellipsoids on the geological map.

The first goal of this section is to take advantage of the available DEMs to provide an accurate map of major drumlin fields of the last ice age and associated ribbed moraines in Switzerland northern Alpine foreland and adjacent areas (France and Germany). The second goal is to analyze their distribution and morphology/morphometry. Morphometry refers to the measurement of morphological parameters that describe the shape and orientation of drumlins. The study of DEMs in a GIS environment is particularly adapted to morphometric analysis because it permits to extract automatically the dimensions of digitized drumlins. Statistical analysis of these data allows to distinguish and compare different drumlin fields or to recognize spatial trends in a given drumlin field.

6.2 Methods

Subglacial landforms were mapped mainly from the DHM25. LiDAR data were used only for Western Switzerland and SRTM data for very small areas beyond the northeastern limit of the DHM25. Landforms were delimited according to their break of slope, recognized visually on the slope attribute, and stored in two polygon shapefiles: drumlins.shp and RibbedMoraines.shp. Drumlin fields were stored in the polygon shapefile drumlin_fields.shp.

For each drumlin, morphometric values were derived automatically for the 2D geometry (length, width, azimuth, area, centroid coordinates) and the elevation statistics derived from the DEM cells within each polygon (minimum, maximum, mean, standard deviation, highest point coordinates). Once extracted, these values were combined to obtain secondary attributes: height, elongation ratio, height/width ratio, volume. Most of these morphometric values were also calculated for ribbed moraines. Table 6-1 presents the methods used to obtain these values.

Two new morphometric parameters were defined to characterize the drumlins: longitudinal and transversal asymmetry (Fig. 6-3). The longitudinal asymmetry is the ratio of the distance between the projections on the drumlin long axis of the highest point and polygon gravity centre, divided by the drumlin length. Its value ranges from -1 to +1. It is positive if the highest point is located up-flow of the gravity center (normal situation for a typical drumlin), negative in the opposite case, and zero if no longitudinal asymmetry is present. The transversal asymmetry follows the same concept, but the distance between the projection of the highest point and the projection of the polygon gravity centre is measured perpendicularly to the long axis and divided by the width. Looking down-ice, the value is positive if the highest point is located left of the gravity center, negative in the opposite situation.

Drumlin statistics were summarized for each drumlin field by spatially joining the drumlins shapefile to the drumlin fields shapefile. For the mean azimuth (ArcToolbox...
Chapter 6 - Drumlins in Switzerland

>> Spatial Statistics >> Linear Directional Mean), the weight of each drumlin is proportional to its length.

6.3 Results

Following this DEM-base analysis, a total of 1237 drumlins and 273 ribbed moraines were accurately mapped in the northern Alpine foreland and in the Jura valleys (Figs. 6-4 and 6-7) in a zone encompassing Switzerland and adjacent areas (France to the west and Germany to the east). Given the limited resolution of the DHM25, the map omits some drumlins smaller than 200 m like, for instance, those of Irchel, west of Lake Zürich.

Landforms were grouped into 22 fields, which are described below in association with three complementary figures. Figure 6-5 presents the general topographical

<table>
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<th>Attribute</th>
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<th>Method</th>
</tr>
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<td>Length (m)</td>
<td>length</td>
<td>For drumlins, creation in ArcView of a separate line shapefile (longaxis.shp) representing the longest straight line connecting two vertices within each polygon (Jenness, 2003). Line direction flipped if necessary in ArcGIS to correspond to ice flow (TTL Corp, 2005). For ribbed moraines, length was calculated in JMicroVision 1.2 (Roduit, 2006) because the “longest straight line” technique was not appropriate to their bended path.</td>
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<td></td>
<td>Width (m)</td>
<td>width</td>
<td>For drumlins, width was computed in ArcMap from the area and length by an approximation of the drumlin shape to an ellipse: width = 4 * area / (PI * length). For ribbed moraines, width was computed with JMicroVision 1.2 (Roduit, 2006).</td>
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<td>Azimuth (°)</td>
<td>azimuth</td>
<td>For drumlins, azimuth was derived from longaxis.shp with ArcMap &gt;&gt; Field Calculator &gt;&gt; polyline_Get_Azimuth_9x.cal (Tchoukanski, 2004). For ribbed moraines, azimuth was computed in JMicroVision 1.2 (Roduit, 2006).</td>
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<td>Calculated with the ArcView extension Polygon centre of mass v1.a (Jenness, 2004a). Results are in the shapefile centerofmass.shp.</td>
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<td>Calculated with the ArcView extension Nearest features (Jenness, 2004b)</td>
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<td>Distance to nearest neighbour (edge to edge)</td>
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<td>Calculated with the ArcView extension Zonal statistics ++ (Beyer, 2006).</td>
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<tr>
<td>Elevation statistics</td>
<td>Mean elevation (m)</td>
<td>z avg</td>
<td>Once the maximum elevation was calculated for each polygon, the drumlins polygon shapefile was converted to a grid (polgrid25) with a cell size equal to the original DEM (dhm25), where grid value inside each polygon corresponded to its maximum elevation. (ArcToolbox &gt;&gt; Conversion Tools &gt;&gt; To Raster &gt;&gt; Feature to Raster). To obtain a grid containing only the values of the drumlins highest points, a new grid (maxv) was created containing polgrid25 values only where polgrid25 was equal to the original DEM (dhm25). This grid was then converted to a point shapefile (maxvalue.shp). This was done in ArcMap &gt;&gt; Spatial Analyst &gt;&gt; Raster Calculator with the following commands: temp = con(dhm25 == polgrid25, polgrid25, -9999) maxv = SetNull(temp == -9999, temp) maxvalue = GridPointShape [maxv] X and Y coordinates were added as new columns in the table of the maxvalue point shapefile (ArcTools &gt;&gt; Data Management Tools &gt;&gt; Features &gt;&gt; Add XY Coordinates). Finally, the coordinates of these points were transferred to the original drumlins polygon shapefile, based on their common location (spatial join in ArcMap).</td>
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Table 6-1 Morphometric parameter calculation for drumlins and ribbed moraines.
setting of each drumlin field, through a blend of DEM elevation and hillshading. Figure 6-6 presents the morphology of individual drumlins and ribbed moraines, through the DEM low-cut elevation. Finally, figure 6-7 presents the landform mapping. In the titles of the following sub-chapters, the letters in parenthesis (a,b,c,...) refer to these three figures. The morphometric values of each drumlin field are summarized in Table 6-2 and plotted in figure 6-11.

6.3.1 Jura ice cap

Three drumlin fields are present in the Upper Orbe Valley area: the Joux, Solliat and Rousses drumlin fields. Situated in a SW-NE oriented synclinorium of the Jura mountains, the Upper Orbe Valley is divided by the French-Swiss border into the Rousses valley to the SW and the Joux valley to the NE, named after their respective lakes fed by the Orbe river. The Upper Orbe Valley is U-shaped with steep flanks and a flat bottom, gently sloping (1.95°) towards the NE, passing from 1060 to 1004 m. The valley is closed at its northeastern extremity and outflow of the Orbe occurs by sub-lacustrine karst (Aubert, 1959). This closed-basin condition led to the accumulation of large volumes of ice during the Würmian glaciation. The Solliat Valley is a smaller parallel valley located along the northwestern side of the Joux valley, with an elevation ranging from 1060 to 1020 m towards the NE. At the Würmian maximum, the Jura ice sheet rose to more than 1800 m north of Geneva (Campy, 1992) and approximately 800 m of slow-moving ice covered the valley. The closed basin geometry of the Joux Valley led to the accumulation of a large volume of dead ice at the end of the glaciation, allowing good preservation of the subglacial features formed after the glacier had become inactive.

Following pages:

Fig. 6-5 Drumlin fields and ribbed moraines in Switzerland - Hillshaded view blended with elevation color scale. a) Joux and Solliat valleys, b) Lake Les Rousses, c) Bornes plateau, d) Bière area, e) Geneva, f) Jegensdorf ribbed moraines, g) Lake Thun, h) Reuss glacier, i) Zürich oberland, j) Lake Constance, k) location map with WGM ice extent. Tick interval: 2000 m. Note that scale is two times larger for a, c and d, and four times larger for b.

Fig. 6-6 Drumlin fields and ribbed moraines in Switzerland - Low-cut filtered DEM. This attribute emphasizes the glacial landforms.

Fig. 6-7 Map of drumlin fields and ribbed moraines in Switzerland. Arrows indicate ice flow.
Chapter 6 - Drumlins in Switzerland

Legend
- drumlin
- ribbed moraine
- till patches
In the **Joux Valley**, drumlins are scattered between the Lake Joux southwestern end and the Praz-Rodet area (502.5/157.0). They have an average length of 330 m and an average width of 96 m and are oriented parallel to the valley axis. They are irregularly shaped but most of the drumlins have a longitudinal asymmetry indicating an ice flow towards the NE (e.g., 507.5/162.0). The drumlins have a low relief and are partly buried by lacustrine and fluvial sediments. Some low-relief drumlins may be completely covered by these younger sediments and invisible today. Between drumlins, post-glacial peat bogs form slightly elevated plateaus. According to the geological maps and field observations, all these drumlins are constituted of till.

The Joux Valley drumlins were interpreted initially by Aubert (1943) as lateral or median moraines. Based on their morphology and distribution, Fiore et al. (2002) reinterpreted them as drumlins, distinguishing till and gravel drumlins. In the gravel drumlins, sharp truncation of the gravel layers at the surface was taken as an argument for a drumlinization of esker deposits during a catastrophic subglacial meltwater flood. As illustrated in section 7.2.3, the new LiDAR DEM shows that the Joux Valley esker was not drumlinized but has preserved its original esker morphology. This invalidates the previous descriptions of gravel drumlins and the meltwater flood erosional phase.

**Solliat Valley (a)**

In the **Solliat Valley**, the drumlins are similar to those in the Joux Valley but occur isolated or grouped, in association with parallel limestone beds, the latter presenting a distinctive sharper morphology. The Solliat drumlins have an exceptionally high height/width ratio (0.18) because many of them contain an elongated and prominent limestone core. However, these beds do not govern the drumlin orientation, as shown by the 10° angular unconformity between the orientation of the limestone beds (50°) and drumlins (60°), north of Le Sentier (506.1/161.8 and 505.2/161.3).

Grouped drumlins correspond to the drumlinization of larger till patches. These till patches do not appear only in the Solliat Valley but are also scattered north and south of the Joux and Solliat valleys, where their smooth surface contrasts clearly with the rugosity of the stratified, faulted and karstified limestone bedrock. The till composition of these patches is confirmed by the geological map (Aubert, 1941). Outside of the Joux and Solliat valleys, these patches correspond to treeless zones, whereas limestone zones are covered by forest. They are generally elongated in the NW-SE direction, perpendicularly to the tectonic trend. Some till patches are located on the lee-side of a topographical step and indicate the ice flow direction (e.g., 513.4/164.6).

At the southwestern end of the Upper Orbe Valley, the **Rousses drumlin field** covers an area of 3 x 0.5 km south of Lake Les Rousses, along the southeastern border of the valley. This drumlin field is particularly well developed and has never been reported previously. Most of the drumlins have a distinct and steep nose at their northern extremity and fade out towards the south, indicating a flow towards the SSW (210°), opposite to the flow indicated by the Joux drumlins. The Rousses drumlins are small (mean size of 65 x 7.4 m) and have a low elongation ratio (between 2 and 3). Their shape is relatively constant and close to the usual drumlin shape. In the southern part, drumlins are parabolic, with a steep and narrow stoss side and a widened lee side. Drumlins are aligned in clusters of W-E to NW-SE orientation interpreted as poorly developed ribbed moraines.

The Rousses drumlins are more regular in shape and distribution than the Solliat and Joux drumlins. This difference may result from different glacial dynamics between the southwestern and northeastern extremities of the Upper Orbe Valley. The Rousses drumlin field is situated at the point where the valley widens, passing from a width of 1500 m at the beginning of the field to 3000 m at the end. This widening led to a transverse extension in the ice, a condition often associated to drumlins formation (Patterson & Hooke, 1995). In addition, the ice flow was unconstrained at the southwestern end of the valley, whereas it was limited to the NE by the dead-end configuration of the valley.

The opposite flow direction of the Joux-Solliat and Rousses drumlins indicates that the Upper Orbe Valley was crossed by the Jura ice cap ice divide, located between the Joux and Rousses drumlin fields. Drumlins could not be generated close to the ice divide because basal velocity was too slow. This information further constrains the WGM reconstitution of the Jura ice cap of Campy (1992).

**6.3.2 Arve glacier (c)**

Formed by the Arve glacier, the **Bornes drumlin field** (Haute-Savoie, France) is located on the Bornes plateau south of Geneva, between the Salève mountain to the NW and the Alpine front to the SE. The numerous and regularly shaped drumlins present a homogeneous direction of ~215° (SW), consistent with the WGM reconstitution of Jäckly (1970) for the Arve glacier. Most of the drumlins form alignments perpendicular or oblique to ice flow, interpreted as minor ribbed moraines. They are concentrated on the highest zone of the plateau at an altitude of ca. 900 m.

**6.3.3 Rhone glacier**

**Geneva drumlins (d)**

LiDAR data confirmed the mapping by Paréjas (1938) of elongated, SW-NE oriented drumlins to the
NW and SW of the Geneva airport, which span a SW-NE elongated area between Ferney-Voltaire (France) and Satigny (Switzerland). These are particularly developed in the southwestern prolongation of the airport, where they show a high elongation ratio (up to 10, mean value of 4.97 (Fig. 6-11e), probably indicative of fast ice flow (Stokes & Clark, 2002). Boreholes indicate that all the drumlins have a till composition. According to the deglaciation model of the Geneva basin of Monjuvent (in Donzeau et al., 1997), the drumlin field area corresponds to the till unit Gy6, deposited during step 6 of the deglaciation, with the Rhone glacier front located in Aire-la-Ville, only 2 km downflow from the most external drumlins. At the same time, the Russin delta formed in a proglacial lake at a level of 425 m. A glacio-lacustrine unit (GLy7) drapes the lowest parts of the drumlin field.

**Apples hummocky ribbed moraines (e)**

Smooth rounded hummocks are present around the locality of Apples (Figs. 6-9 and 7-19). They lie on the northeastern part of a SW-NE elongated topographic shoulder bordered by the Graviers de la Côte escarpment. Previous authors called these hills drumlins (Custer & Aubert, 1935) or buttes de la région des bois (Vernet, 1973b). They are the upper surface of a thick unit of clay-rich till with striated gravels or boulders (Vernet, 1973a). The Rhône glacier deposited this till subglacially during the Würm, directly over the Lower Freshwater Molasse (Chattien). Based on the till lithology, former authors have attributed a subglacial origin to these landforms but did not provide an interpretation of their particular shape and their origin remains unknown.

The smooth and poorly oriented relief of the Apples hummocks contrasts with the sharp and oriented Molasse rock drumlins to the east. Based on their planform, they can be subdivided into two categories: round hummocks with a round or slightly elliptical planform and elongated hummocks oriented W-E (Fig. 7-19). Both forms are 10-15 m...
high and ~ 400 m wide. Their original morphology has been altered in places by late- or post-glacial fluvial erosion.

**Hummocks, hummocky landscapes or hummocky moraines** have often been reported in areas formerly covered by an ice sheet, encompassing a wide range of morphologies and different mode of formation involving supraglacial, proglacial or subglacial processes. However, many authors keep the term hummocky moraine to describe chaotic landscapes resulting from the differential accumulation of the supra-glacial debris cover and subsequent melting (see Benn & Evans, 1998, p. 483-487 for a review). Such hummocky landscapes are more chaotic and uneven than the Apples hummocks and do not constitute a satisfying equivalent.

The Apples elongated hummocks are much more similar to **ribbed moraines** (see section 2.2.1). Hättestrand (1997) proposed a geomorphological classification of ribbed moraines into four types (Fig. 6-8). One of these types is called **hummocky ribbed moraine** (HRM) and is described as “a poorly developed form of the Rogen moraine, where the ridges are less constant in height and spacing. The ridges are clearly transverse to ice flow, although they are not as strictly parallel as within Rogen moraines. Their size is similar to Rogen moraine ridges, but they are commonly shorter in length.” This description and the sketch presented in figure 6-8 clearly match the Apples hummocks.

Besides these elongated hummocks, numerous round hills are present in the Apples area, which are not mentioned in the above description of hummocky ribbed moraines, or in other publications dedicated to ribbed moraines. However, we could find several examples of round hummocks on DEMs or satellite images, in close association with ribbed moraines (Fig. 6-9). Round hummocks may occur at the outer margin of the ribbed moraines area (Fig. 6-9b,c), at the transition between drumlins and ribbed moraines (Fig. 6-9c), or at the limit between two sets of ribbed moraines with different orientations (Fig. 6-9d). These three examples show that round hummocks are common within hummocky ribbed moraines and that their width is similar (~500 m) despite the different size of the Alpine, Laurentide and Irish ice sheets. Little attention has been paid to these round hummocks in the literature, probably because of their scarcity and common morphology compared to associated drumlins and ribbed moraines.

What conditions did lead to the formation of the Apples HRM and rounded hills? As visible on the Quaternary isopach map (Fig. 4-7), the Arzier - L’Isles plateau is an isolated accumulation of glacial deposits preserved from glacial erosion. This preservation results from the particular position of the area relative to the WGM ice flow. The Rhone glacier northeastern branch flowed through the broad valley linking the Morges-Lausanne zone to Yverdon and Lake Neuchatel, while the southwestern branch followed the Lake Geneva depression. The Arzier-L’Isles plateau corresponded therefore to a zone of diffuence between the northeastern and southwestern branches of the Rhone glacier. The general topography, end moraine disposition and Molasse drumlin orientation indicate that the Apples area is not located exactly at the diffuence point but slightly more to the northeast. This intermediate position could explain the association of elongated transverse bedforms such as HRM with rounded hummocks, corresponding respectively to an established or variable ice flow direction.

**Jegenstorf ribbed moraines**

Because of the near total absence of glacial material over the Molasse bedrock, the Solothurn arm of the Rhone glacier did not produce till drumlins but rock drumlins, presented in chapter 5. However, a small set of low relief, discontinuous ridges oriented SSW-NNE is present in the...
5 km southwest of Annecy, west of the Semnoz

Around the locality of Montmélian, 15 km south of
Southeastern part of the Isle Crémieu area, 45 km east

Jugenstorf area, near the confluence of the Rhone and Aar glaciers. Pugin (1989) reported glaciofluvial gravels within these ridges. We interpret them as ribbed moraines formed by the Rhone glacier. They are not associated with drumlins.

Other drumlins associated with the Rhone glacier

While the glaciers of eastern Switzerland produced numerous drumlins, the Lyon arm of the Rhone glacier did not produce significant drumlin fields on the Swiss territory. Examination of SRTM data suggests that the Rhone glacier could have formed drumlins further down-flow, beyond the Swiss-French border. Despite the relatively low resolution of SRTM data, we identified several potential drumlin fields:

- 5 km southwest of Annecy, west of the Semnoz mountain range
- around the locality of Bellev, 10 km west of Lake Bourget
- around the locality of Montmélian, 15 km south of Lake Bourget
- Southeastern part of the Isle Crémieu area, 45 km east of Lyon.

Beyond the limits of the WGM, the Dombes area also presents interesting glacial landforms, though associated to the most extensive glaciation (MEG, equivalent to the Russian). Situated to the NE of Lyon, the Dombes area is a slightly undulating plateau of circular shape with a diameter of ~40 km. Below a superficial cover of loess, the plateau is constituted of clayey till (Moiriat, 2003). It is bounded to the west and the north by the moraines of the MEG. Numerous elongated glacial hills and more than 1100 ponds form a fan-like pattern parallel to the former ice flow (Fig. 6-10). Created artificially since the 13th century, the ponds accentuate the glacial landscape, which corresponds probably to drumlins or flutings.

6.3.4 Aar glacier (g)

The Thun drumlin field is located in the Aar Valley, west of Thun, on a highland bounded to the west by the Gürne River and to the east by the Aar River. Drumlins form a radial pattern, their orientation ranging from 250° in the south to 345° in the north. Most of the drumlins are superimposed over prominent transverse ridges. The latter were interpreted as fronto-lateral moraines (Beck & Rutsch, 1949) or intermediate moraines (Wagner, 2001). We reinterpret them as ribbed moraines, because their association with drumlins and their typical shape, which is wider and lower than usual Alpine moraines. These ribbed moraines are perpendicular or slightly oblique to the drumlin long axis. Both landforms are composed of till attributed to the Aar glacier Jaberg stage (late Würm), during which the ice front would have been situated only 5 km down-ice from the drumlin field (Beck & Rutsch, 1949). The radial pattern formed by the drumlin long axis indicates the probable vicinity of the ice front. If they had formed during the WGM when the ice front was much further downflow, all the drumlins would have been parallel to the valley axis.

6.3.5 Reuss glacier (h)

Three drumlin fields were formed by the Reuss glacier in the Zoug area and are centered on the localities of Eschenbach, Sins and Knonau. All the drumlins are oriented towards the north. The Eschenbach and Sins fields are not very rich in drumlins, as shown by their large “nearest neighbour” values (397 and 267). In the Knonau drumlin field, some are aligned en échelon, forming a NW-SE ridge oblique to the ice flow. Unlike the Thun ribbed moraines, this ridge is isolated and corresponds probably to an older fronto-lateral moraine reshaped during the drumlin formation.
6.3.6 Linth glacier (i)

In the valley southeast of lake Zürich, the Glattal-arm of the Linth glacier produced two drumlin fields: Gossau (upflow) and Illnau-Effretikon (downflow). They are bounded laterally by two tunnel valleys, each one being occupied by a lake: Lake Greifen (SW) and Lake Pfäffikon (NE). In both fields, drumlins have a NW-SE direction. The Gossau drumlin field present well-developed drumlins which conducted the Swiss Confederation to classify the area as a geotope (Zürcher Oberland geotope), a zone deserving protection because of its geological interest. In
Chapter 6 - Drumlins in Switzerland

the Illnau-Effretikon field, drumlins are less elongated, especially in the western part. The absence of drumlins between the two fields is probably related to the absence of glacial material over the Molasse bedrock. According to the geological map of Hantke (1967), these drumlins are composed of late Würm till and gravels. The Langfur gravel pit in Gossau (ZH) reveals large gravel foresets deposited during an older glacial readvance (Wildermuth et al., 1982). This gravel sequence has been eroded during the last glacial readvance to form the present drumlins.

6.3.7 Rhine glacier (j)

Around Lake Constance (Bodensee in German), the numerous drumlins form a spectacular radial pattern straddling the Swiss-German border, reflecting the flow lines of the large piedmont lobe of the Rhine glacier. In the southwestern part, the strong rotation of the flowlines towards the left indicates that the large volume of the WGM Rhine piedmont lobe forced the incoming ice to flow laterally. To the east, the Lindau drumlin field is characterized by the presence of numerous ribbed moraines bounded laterally by strongly diverging drumlins (300° for the western drumlins, 30° for the eastern ones). The eastward transition from ribbed moraines to drumlins is characterized by the superimposition of drumlins over the ribbed moraines and the fragmentation of the ribbed moraines into "en échelon" alignments of broad crescent-shaped landforms pointing up-ice. These latter have an intermediate morphology between drumlins and ribbed moraines and may be described either as parabolic drumlins (Shaw, 1983) or as Blatnick moraines (Hättestrand, 1997). The Bodanrück drumlin field, located north of Lake Constance, presents en échelon drumlin alignments and transverse asymmetrical drumlins (Shaw, 1983) in its southwestern border. These drumlins result from the erosion of a Würm sequence starting with fine-grained till and lake deposits, followed by proglacial gravels and ending up with an upper till, often deformed and sheared (Ellwanger, 1992). Gravels and upper till were deposited during the last major ice readvance (Innere Jungendmoräne).

6.4 Discussion and conclusions

6.4.1 Drumlins

Drumlin dimensions

Compared to other drumlins worldwide (Fig. 6-12), drumlins of Switzerland are of medium to small size. Despite the relatively small size of the Alpine ice cap, the largest Alpine drumlins and ribbed moraines present shapes and dimensions very similar to those found under the larger North-American ice sheet (Fig. 6-14). The drumlin dimensions (length, width and height) show a large standard deviation within each drumlin field. This variability is not related to a down-ice trend, as it is the case in some other drumlin fields (e.g., length increase towards the former ice margin, Evans, 1987).

In a given drumlin field, width is more constant than length and permits the classification of drumlin fields into three categories (Fig. 6-11b): small drumlins (Bornes,
Chapter 6 - Drumlins in Switzerland

Geneva, Jura drumlins), intermediate drumlins (Thun) and large drumlins (other fields). Small drumlins were recognized only in Western Switzerland thanks to the high-resolution LiDAR DEM, but their presence elsewhere in Switzerland is not excluded: the map of Früh (Fig. 6-1) indicates drumlin fields invisible on the DHM25, such as the small field lying 5 km west of Effretikon. Examination of Lidar data would determine if the size of the landforms composing these fields are below the resolution of the DHM25, or if the map of Früh is erroneous.

The mean elongation ratio is generally situated between 2 and 3, but reaches an average value of 5 in the Geneva drumlin field. The average height/width ratio ranges from 0.07 to 0.12 but is higher for the Solliat drumlins (0.185) because most of them are built on top of prominent limestone highs.

For almost all drumlin fields, the average longitudinal asymmetry is slightly positive (between 0 and 0.1), indicating a steeper stoss-side and a gentler lee-side in agreement with the typical drumlin morphology. However, this asymmetry is not very pronounced and many drumlins present a negative asymmetry (lee-side steeper than stoss-side). Transversal asymmetry depends mainly on the slope of the drumlin field transverse to drumlin orientation. The Solliat drumlins present a particularly strong positive transversal asymmetry of 0.29 (highest point to the left of the drumlin long axis), because they are located on the slope of the valley’s southeastern flank.

The plotting of drumlins azimuth on a Rose diagram (Fig. 6-13) shows two families of orientation reflecting the ice flow and indirectly the orientation of the Alpine valleys: SW-NE (parallel to the Alpine range) and NW-SE (perpendicular to the Alpine range).

**Drumlin distribution**

Drumlins are concentrated on the northern Alpine foreland and in the Jura valleys. They seem to be absent on the southern side of the Alps, but may be hidden by post-glacial alluvial sediments. They occur near the ice margin but do not reach the external moraines. They are generally bounded laterally by tunnel valleys.

Drumlins are absent in the inner Swiss Alps. However, drumlinoids (i.e., bedforms similar to drumlins, but with a less typical shape) have been previously reported at the front of present valley glaciers of Switzerland (van der Meer & van Tatenhove, 1992). By contrast with drumlins, drumlinoids are rare and small, and occur isolated or in small groups. They were not recognized on the DHM25.

No clear relation exists between the regional large-scale topography and the occurrence of drumlins. Most of the latter occur in lowlands or on valley sides, but the Bornes drumlins are concentrated on a slightly elevated plateau. Generally the drumlin field longitudinal profile is horizontal or slightly ascending (e.g., 1.3 % for the Lindau drumlin field), but it may also be slightly descending (e.g., Muellheim drumlin field). The absence of relationship between topography and drumlin occurrence confirms the work of Patterson and Hooke (1995).

The distribution of drumlins reflects primarily the distribution of glacial material because drumlins do not form directly on bedrock. This is particularly visible in

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**Fig. 6-14** Comparison of the drumlins and ribbed moraines under the Rhine piedmont glacier (Alpine ice sheet) and under the Green Bay lobe (Wisconsin, Laurentide ice sheet). Note the strong similarity in size and shape, despite the different size of the two ice sheets.
the Rhine glacier piedmont lobe, where bedrock bumps are not drumlinized.

Several authors have noted a diverging pattern in drumlin fields and linked the presence of drumlins to diverging flow lines and extensive strain transverse to ice flow (e.g., Patterson & Hooke, 1995). In Switzerland, only the Rhine and Aar glaciers present a radial pattern. From the absence of radial pattern in other fields, we conclude that an extensive transverse strain is not a prerequisite to drumlin formation.

6.4.2 Ribbed moraines

Six groups of ribbed moraines were recognized in the investigated area (Table 6-3). They constitute the first report of ribbed moraines associated to the Alpine ice cap. Alpine ribbed moraines have the same morphological characteristics and dimensions as other ribbed moraines worldwide (Table 6-3). The presence of ribbed moraines under the relatively small Alpine ice cap increases the range of physical conditions compatible with their formation.

<table>
<thead>
<tr>
<th>RM field</th>
<th>Glacier</th>
<th>RM width (m)</th>
<th>RM spacing crest to crest (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rousses</td>
<td>Jura ice cap</td>
<td>100-230</td>
<td>180-300</td>
</tr>
<tr>
<td>Bornes</td>
<td>Arve</td>
<td>250-300</td>
<td>300-350</td>
</tr>
<tr>
<td>Apples</td>
<td>Rhone</td>
<td>230-400</td>
<td>300-500</td>
</tr>
<tr>
<td>Jegenstorf</td>
<td>Rhone</td>
<td>300-500</td>
<td>400-600</td>
</tr>
<tr>
<td>Thun</td>
<td>Aar</td>
<td>250-400</td>
<td>250-400</td>
</tr>
<tr>
<td>Lindau</td>
<td>Rhine</td>
<td>300-600</td>
<td>350-1000</td>
</tr>
</tbody>
</table>

Table 6-3 Dimensions of ribbed moraines (RM) recognized in this study.

6.4.3 Formation

Formation time

For the Rhine, Linth and Reuss glaciers (Hantke, 1978; Keller & Krayss, 2005), drumlins are located between the Stein-am-Rhein (19.5 cal kyr BP, also known as Innere Jung-Endmoräne) and Konstanz readvances (18 cal kyr BP). Drumlins are absent between the WGM stadium and Stein-am-Rhein readvances, a situation already reported in the German Alpine foreland (Habbe, 1989; Habbe, 1992). The only exception is the Illnau-Effretikon drumlin field.

In the large Quaternary ice sheets, the relation between end-moraines and drumlin fields suggests that drumlins formed time-transgressively as the ice front retreated, each recessional phase being characterized by a narrow drumlin-forming zone located a few kilometers back from the margin (e.g., Goldstein, 1994; Hättestrand, 1997; Clarhäll & Jansson, 2003).

In the case of a time-transgressive formation during ice retreat, the drumlins would have an age ranging from 19.5 to 18 cal kyr BP. The Illnau-Effretikon drumlin field formed earlier, between the WGM (23 cal kyr BP) and the Stein-am-Rhein stadium. However, it must be noted that drumlins are generally not formed in the outermost part of the ice sheets, where ice is thinner (Aario, 1977a). Therefore, drumlins in the Alpine foreland drumlins could have been formed during the WGM, in spite of their absence near the ice front.

These considerations do not concern the Rhone glacier, where well-developed drumlin fields were not recognized on DEMs.

Processes

In Switzerland, the genesis of drumlins has been little investigated. It was suggested recently that the majority of the elongated hills mapped as drumlins on geological maps were in fact intermediate moraines resulting from the Alpine icefield dendritic pattern (Wagner, 2001; Wagner, 2002; Wagner, 2003; Hantke & Wagner, 2004). Our observations go against this radical reinterpretation, already refuted by Graf et al. (2003).

To explain the absence of drumlins between the Stein-am-Rhein readvance and the WGM, Habbe (1989; 1992) argued that drumlin formation in the Alpine foreland required a glacier readvance over a ‘decaying permafrost’, where the irregularities at the surface of the sinking permafrost table provided nuclei for drumlin formation by differential erosion. The permafrost conditions are supported by the lack of glacial deformation in interglacial clayey sediments. These conditions would have occurred only during the IJE readvance because the ground was frozen during older readvances and completely defrozen during the younger ones, preventing drumlin formation. Ellwanger (1992) proposed the same interpretation for the Bodanrück drumlin field (Rhone glacier), based on sedimentary and glacio-tectonic evidences. A similar scenario has been proposed for the southern Laurentide ice sheet (Stanford & Mickelson, 1985; Cutler et al., 2000).

Our study evidences the presence of ribbed moraines in association with drumlins. The drumlinization of ribbed moraines and the progressive spatial transition between the two landforms suggests a synchronous formation and a strong genetic relationship. Any formation theory should thus provide an explanation for both landforms and take into account the field evidences supporting an erosive origin.

The “permafrost nuclei” hypothesis (Habbe, 1989; Ellwanger, 1992; Habbe, 1992) is unlikely, as it cannot account for the formation of ribbed moraines. Hypotheses of ribbed moraine formation as glaciotectonic products (see section 2.2.1) are rejected because they do not take into account the genetic relationship with associated drumlins. Finally, a formation of ribbed moraines resulting from drumlins reorientation by subglacial deformation after a
change in ice flow direction (e.g., Boulton, 1987) is also rejected, because the Alpine ice flow was constrained by topography.

In the case of a formation by catastrophic subglacial meltwater floods (Fisher & Shaw, 1992; Shaw, 1998), the synchronicity of drumlin formation time in disconnected valleys would require the occurrence at the same period of at least one contemporaneous huge jökulhaup in each valley. Given the random occurrence of these events, such a widespread jökulhaup period is unlikely.
Chapter 7
Glacio-fluvial Deposits and Moraines in Western Switzerland

7.1 Introduction

This chapter investigates the geomorphology of glacio-fluvial deposits and moraines in Western Switzerland, based on the Geneva and Vaud high-resolution LiDAR DEMs and on new sedimentological observations. Numerous new landforms are recognized and the morphological characteristics of known landforms is refined. Four areas have been studied in detail: Bois de Chênes, Bière, Joux Valley and the Lake Geneva shore between Lausanne in Geneva.

7.2 Glacio-fluvial deposits

7.2.1 Bois de Chênes area

The Bois de Chênes (BdC) is a small natural reserve located 6 km north of Nyon in the commune of Coinsins, in the foothills of the Jura mountains and at the southwestern end of the Graviers-de-la-Côte formation (Arn, 1984). This zone presents a network of subparallel, SW-NE oriented ridges composed of stratified gravel and sands. Further westwards, the LiDAR DEM also reveals two other sets of ridges in the communes of Trélex and Gingins presenting a morphology and orientation similar to those of the BdC ridges. We will refer to these three zones (BdC, Trélex, Gingins) as zone 1, zone 2 and zone 3. In the three zones, the ridges lie on a late-Würm till unit, which itself lies on the Graviers-de-la-Côte (middle Würm) in zone 1 (Arn, 1984; Weidmann & Arn, 2005).

Geomorphology

Because a large part of the BdC ridges area is covered by forest, field interpretation is difficult and the detailed, vegetation-free Lidar DEM is particularly appropriate (Figs. 7-2 and 7-3). Zone 1 extends over an area of 2 x 1.5 km with altitudes ranging from 590 m in the north to 450 m in the south. It is bounded to the west and east by the Montant and Serine rivers, although an isolated ridge is also present on the left shore of the Serine River. Individual ridges are symmetrical in cross-section, 3-10 m high, 20-60 m wide, and form an anastomosing pattern with a ridge spacing of 30-300 m. Where two ridges form a closed depression, ponds or marshes occupy the latter. The infill of these lows by deposits younger than the ridges may form small elongated, flat-topped plateaus. Some ridges have a sinusoidal planform, with a wavelength of 100 m and an amplitude of 30 m. Although these ridges were principally studied in the BdC by previous authors, ridges are also present south of the BdC, around the locality of Coinsins. In the southern part, gravel quarrying and, to a lesser extent, agricultural activity have altered the ridges morphology, sometimes completely removing some of them. North of zone 1, a terrace presents a height (590-560 m) and slope (3.4 %) similar to those of the adjacent ridges.

In zone 2, sharp ridges are located in the northern part (places called Mollard Parrelliet, Bois de Ban and Bois à la Dame). They also show a SW-NE orientation. The main accumulation consists of a dense anastomosed network. Much smoother ridges are present in the southern part. Like in the BdC, a terrace with height (645-625 m) and slope (5.7%) similar to those of the adjacent ridges seems to overlie the ridges. Two ridges are also present north of this terrace, with N-S and NW-SE orientations.

In zone 3, ridges are located NW of Gingins, over the sites called Château-Blanc and Le Pontet. Two long parallel ridges are visible, with numerous short segments around them.

Further westwards, an elongated, isolated moraine constituted of till stretches along the Jura foothill between Chéserex and Divonne (France). It is aligned with the ridges described above, although it is wider, straighter, lower, smoother and more continuous. The difference in lithology and morphology excludes a direct relation with the ridges of zone 3.

Between zones 1 and 2, sharp ridges are absent over 2.5 km but the ridges of zone 1 seem to extend as much smoother ridges in the southern part of the fan. The latter are probably gravel ridges deposited at the bottom of a
broad and low depression and later covered by alluvions from the rivers “La Colline” and “Le Montant”, located to the west and the east.

The zone separating zone 2 from zone 3 is occupied by a low relief, 600 m wide alluvial fan formed by a temporary river located above in the “Creux Mariot” Valley. The alluvial activity eroded the ridges and buried their possible remnants.

Sedimentology

Gravels and sands in zone 1 are well rounded, not striated, and come from the Jura (2/3) and the Alps (1/3) (Aeberhardt, 1901). They are perfectly stratified with some discordant stratifications (Fig. 7-5) and form overall a pseudo-anticlinal structure concordant with the ridge topography (Aeberhardt, 1901). Grain size is coarser at the centre of the ridges and decreases towards the sides (Nguyễn-Tang, 2002).

In zone 2, ridges are also formed of stratified gravel and sands (Aeberhardt, 1901) and their exploitation has led to a few abandoned gravel pits affecting their original morphology. A small fresh outcrop was encountered on the crest of a NNW-SSE oriented ridge north of the terrace (503.630/142.670). Sands and gravels plunge with a dip of 240/25 perpendicularly to the crest orientation.

We have not found any fresh outcrop in zone 3 but the disseminated gravels and boulders found in three abandoned pits indicate that these ridges are constituted of gravels and sands. The most recently abandoned gravel pit is encountered at the intersection of the route de la Dôle and the chemin du Creux (502.5 / 140.95). Following the
Fig. 7-2 Hillshaded view of the Bois-de-Chênes ridges and geomorphological map. See Fig. 7-1 for location.

Fig. 7-3 LiDAR elevation profile A-A' perpendicular to the Bois-de-Chêne ridges. See Fig. 7-2 for location.
route de la Dôle uphill, two other abandoned gravel pits (502.48 / 141.12; 501.9 / 141.1) are covered by vegetation but show small outcrops of light-coloured diamict at the top of the excavation front. Underneath the vertical diamict outcrop, the excavation front has collapsed and is covered by vegetation. These collapsed deposits probably correspond to non-cohesive sands and gravels, more easily eroded than the overlying till.

Origin of the deposits

The particular geometry of the BdC deposits has drawn the attention of many authors who proposed different interpretations.

Schardt (1898) and later Arn (1984) considered them as frontal moraines built up during the readvance of the Jura glaciers after the retreat of the Rhône glacier. Aeberhardt (1901) interpreted them as subglacial meltwater channels (eskers) modified by ice movement, formed below the Rhone glacier, during its retreat. This point of view was also followed by Nguyên-Tang (2002), who considered the grain size decrease away from the crest as a typical characteristic of deposition in a subglacial meltwater channel, in which water flow is stronger at the centre; the esker sedimentation would have been followed by a catastrophic meltwater flood leading to the drumlinization. Finally, Weidmann and Arn (2005) focused their attention on the depressions between the ridges, interpreting them as dead-ice collapse structures within periglacial alluvions.

None of these authors genetically linked the three ridge zones and their interpretations focuses only on zone 1. However, Aeberhardt (1901) noted that the shape and orientation of the ridges in zones 2 and 3 resemble the ridges of zone 1. He interpreted them as lateral moraines of the Rhone glacier, connected to the long and smooth morainic ridge stretching along the Jura foothill between Chésérex and Divonne (France).

The ridge morphology is clearly different from that of known moraines in Western Switzerland such as the parallel Montosset moraines (section 7.3.2), the wide and smooth Chésérex-Divonne moraine or the numerous other Jura foothill moraines. Besides, Jura glaciers would have built arched, convex downslope moraines, which is not the case.

When searching in the scientific literature for similar ridges interpreted as moraines, two similar sets of ridges can be encountered. They are located in the Isortoq valley (Fig. 7-6 a) in north-central Baffin Island of Canada (Andrews, 1963; Ives & Andrews, 1963) and in the Swedish mountains (Fig. 7-6b, Borgström, 1979). Both examples are De Geer moraines (called cross-valley moraines in Baffin Island), i.e., sub-parallel moraine ridges with a saw-tooth pattern formed in a glaciolacustrine setting. Despite their apparent similarity with the BdC ridges, De Geer moraines in Baffin Island and Sweden present significant differences. First, they are constituted of till or glaciotectonized material, not of stratified gravel and sands. Secondly, they present a clearly asymmetrical cross-section with a gentle proximal side and a steep distal side, whereas the BdC ridges are symmetrical. Finally, the formation of De Geer moraines requires a glaciolacustrine environment, which is unlikely in the BdC area, because no glaciolacustrine sediments were observed. Consequently, the BdC ridges cannot be assimilated to De Geer moraines and the moraine interpretation is rejected.

With respect to the esker hypothesis, numerous similar examples have been described worldwide and named multiple-crested eskers (Shreve, 1985),
The presence of N-S oriented ridges in the northern part of zones 1 and 2, associated to sublinear ridge orientation between zones 2 and 3 leads to the grouping of the ridges into two main esker networks: zone 1 and zone 2-3. Meltwater driven by gravity came down the steep Jura slopes down to the northeastern limit of zones 1 and 2. From this limit, the slope decreases suddenly and the gravity force became insignificant with respect to the glacio-static pressure. The meltwater path was not driven by gravity anymore but by the subglacial water pressure gradient. It flowed along the Jura foothill towards the southwest to reach the low-pressure zones at the ice margin. This rapid transition between the influence of gravity and ice pressure explains the change in direction of the ridges from south to southwest.

Terraces located in the northern part of zones 1 and 2 are are found in several other places along the Jura foothills and were interpreted by Arn (1984) as kame terraces formed between the Jura and Rhone ice fronts (terrasse adventives de confluence or TAC) during deglaciation. During the formation of these TACs, the esker ridges already liberated from the ice were buried by the TAC material. Therefore, the flat terraces are not genetically related to the eskers.

7.2.2 Ballens glaciofluvial complex

The Bière area is rich in geomorphological features: glaciofluvial accumulations and recessional moraines overlay the background morphology of Hummocky ribbed moraines. These landforms are eroded in places by late-glacial and post-glacial streams. In this section, the Lidar DEM is used to accurately describe the morphology of these landforms. Thanks to new sedimentological evidences, an attempt is made to unravel their formation process.

The Bière area is part of a plateau along the Jura foothill, stretching from Arzier to the southwest to L’Isles to the northeast (Fig. 7-1). This plateau is limited to the southeast through an escarpment of ~ 250 m, which fades out towards the northeast. The latter is formed mainly of Alpine glaciofluvial deposits known as la Côte gravels (graviers de la Côte, Arn, 1984). This material is deeply eroded by post-glacial streams and lies on the Chattian Molasse (Vernet, 1973a). The limit between the two units is easily recognizable on the LiDAR NE-hillshaded view thanks to the clear contrast between the eroded gravels and the more uniform Molasse relief (Fig. 7-1). From west to east, the contact rises from 540 to 650 m.

South of Ballens, the topography changes radically from smooth to irregular (Fig. 7-8). This change corresponds to an accumulation of glaciofluvial gravel and sands lying directly on top of the Apples hummocky ribbed moraines. The gravel accumulation is cut by two flat-bottomed abandoned valleys, occupied during ice retreat by periglacial rivers (Arn, 1984). Le Paudex - the westernmost and shortest valley – has a mean height of

anastomosing eskers (Shaw & Kvill, 1984; Shaw et al., 1989) (Fig. 7-7), braided eskers or networked eskers. This geomorphological evidence, together with the previously published sedimentological evidences, confirms the esker interpretation of Aeberhardt (1901) and Nguyen-Tang (2002). However, the drumlinization of the esker deposits proposed by the latter author is refuted because of the absence of streamlining.

The interpretation of the ridges as dead ice collapse structures within periglacial alluvions (Weidmann & Arn, 2005) cannot explain the general ridges morphology. This process may have formed only small surface depressions or sedimentary dead-ice structures resulting from blocks of overlying ice sometimes falling into the esker sediments.

Because eskers are only preserved under inactive ice, this network must have formed at the end of the last glaciation. Morphology and sedimentary structure of the ridges are well preserved, indicating that the esker system was not developed supra- or englacial but subglacially.

These eskers formation at the limit of the Rhone and Jura ice caps indicates that ice was thinner there, leading to the convergence of the Rhone and Jura meltwaters. In large ice sheets, the same convergence occurs at the limit of ice lobes (e.g., Warren & Ashley, 1994; Russell & Arnott, 2003).

The ridge sediments were thought first to come from the erosion of the “Graviers de la Côte” gravel accumulation, to the east. However, these latter do not have the same petrographical composition as the ridges: they consist almost exclusively of Alpine elements (Arn, 1984), while only one third of the ridges gravels come from the Alps, the remaining two thirds coming from the Jura (Aeberhardt, 1901). Moreover, the northernmost ridges of zone 2, and to a lesser extent of zone 1, show a N-S orientation, suggesting a meltwater coming from the Jura. The Rhone elements came probably from the reworking of till deposited by the Rhone glacier on the Jura flanks during the WGM.
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**Fig. 7-8** Ballens glaciofluvial complex, as visible on the LiDAR DEM. See Fig. 7-1 for location.

**Fig. 7-9** Schematic and interpretative section of the Ballens gravels, modified after Arn (1984).
The local geomorphological facies is rather chaotic. which is characterized by a steep downward step of ~ 8 m. The local geomorphological facies is rather chaotic.

Arn (1984) distinguished three gravel units based on geomorphology and stratigraphy. These units are easily recognizable on the Lidar DEM, from west to east (Fig. 7-8): (1) the Ballens gravels, forming a plateau gently dipping towards the east; (2) the Mormontan gravels, an outwash plain associated to the Paudex paleo-river; (3) the Montosset gravels, associated to the paleo-Serine and overridden by the Montosset till deposited during a late-glacial readvance. Arn interprets these three gravel accumulations as proglacial meltwater deposits corresponding to three successive stages of the receding Rhone glacier. Does the Lidar DEM morphologies confirm this interpretation?

The Montosset gravel unit, recognized by Arn (1984) from gravel outcrops, is covered by the Montosset till and therefore lacks a morphological expression. Concerning the Mormontan gravel unit, the DEM hillshaded view shows an association of subaerial channels and interfluvies forming a braided pattern, supporting the interpretation of Arn of a proglacial outwash plain or sandur (Fig. 7-8). The general slope indicates a flow towards the north. Another gravel accumulation located south of Gimel (514.8/151.0, Fig. 7-1) has the same height as the Mormontan unit. Both accumulations were probably formed by the same fluvial system and later eroded by the paleo-Serine.

The Ballens glaciofluvial complex (BGC) does not show any indication of former subaerial streams. It differs clearly from the Mormontan unit through its general geometry and local geomorphological facies. The BGC forms an N-S elongated body of 2.5 x 1.7 km. Starting from the west, it begins with a zone constituted of small glaciofluvial patches distributed over the hummocky till landscape, where the mean elevation is around 700 m. The glaciofluvial nature of these small landforms is confirmed by boreholes and electrical soundings (Holzmann, 1999). The patches form small, SW-NE oriented, esker-like ridges or buttes, 60-70 m wide. To the east of this patchy zone, the elevation rises rapidly, this step marking the western limit of the Ballens gravel zone recognized by Arn (1984). This elevated zone shows a strong west-east asymmetry: along the western limit, a N-S oriented ridge culminates around 725 m, separating the short and steep (~ 5°) western flank from the gently dipping eastern flank (~ 1°). The BGC eastern limit was eroded by the Paudex paleo-river, except its northern part, which is characterized by a steep downward step of ~ 8 m. The local geomorphological facies is rather chaotic.

The eastern flank is dissected by a dendritic network of thin, shallow grooves oriented generally NNW-SSE or NW-SE, corresponding probably to lateglacial stream activity. These could be small streams flowing towards the Paudex valley. This subtle relief is better preserved in the forested zone, unaffected by agriculture.

**Sedimentology**

Boreholes and geo-electrical investigations indicate a mean gravel thickness of 25 m in the area. This corresponds to a total volume of 33 millions m$^3$ of gravels, covered by less than 1 m of soil (DTPAT-SG, 1991). These data also show that the hilly relief of the Apples hummocky ribbed moraines is still present below the gravels and that the deposition of the glaciofluvial material was not preceded by a local erosional event (Holzmann, 1999). The gravel come almost exclusively (90-98 %) from the Rhone glacier (Arn, 1984).

Two gravel pits are located in the BGC (Fig. 7-8). Both present sub-horizontal stratification and lack deltaic stratifications (Arn, 1984). The first pit, located to the north of the gravel accumulation (517.95/155.9), is abandoned and lacks fresh outcrops. Arn (1984, p. 273) made a sedimentological description and reported the following succession from top to bottom:

- 2-3 m of horizontally bedded sands with a few pebbles.
- 6-7 m of medium sandy gravels with sub-horizontal stratifications and channel structures

The second pit, called Les Mossières, is located to the south of the gravel complex, 2.5 km east of Bière. The exploitation has significantly altered the landscape (Fig. 7-10). The pit was briefly described by Arn (1984), who reported at the location 517.5/154.4 the following succession from top to bottom:

- 1-2 m of horizontal sands
- 2-3 m of fine gravels, locally affected by dead ice melting
- ~~~ minor erosional unconformity
- 3 m of horizontal sand with current ripples at the top
- ~~~ erosional unconformity underlined by striated pebbles
- 6-7 m of fine to medium gravels, with sub-horizontal or oblique channel stratifications

The Mossières pit was abandoned after 1984 and has been reactivated in 1999. Today (January 2006), the pit exhibits several spectacular outcrops (Fig. 7-10). Outcrop 1 (Fig. 7-11) is the largest one (10 m high) and reaches the deepest deposits. Two main formations can be identified. The lower formation consists in seven units of poorly sorted, coarsely stratified gravels bounded by sharp erosional unconformities. The apparent dip varies from one to the other, ranging from 25° NE to 25° SW. These sediments indicate important and rapid variations in flow power and direction, typical of subglacial glaciofluvial sedimentation (e.g., Brennand & Shaw, 1996; Fiore et al., 2002; Mäkinen, 2003). This lower formation is only visible
here because the other outcrops in the Mossières pit are not deep enough to reach it.

The second formation has a smaller grain size, consisting essentially of coarse to fine sands. Its boundary with the underlying formation is marked by a main discordance dipping towards the SW, so that the sandy formation is thicker and better exposed in the southwestern part of the outcrop. Similarly to underlying gravel units, the sand beds are generally disconformably overlying each other. Sands present cross-stratifications, sub-parallel stratifications or form massive units. This upper formation can be traced in all the other outcrops, which show a mean flow towards the NW. Downflow, one can observe a decrease in granulometry, a thinning of the units and a transition from discontinuous (Fig. 7-12a) to continuous beds (Fig. 7-12b,c). Outcrop 3 displays the following succession from top to bottom (Fig. 7-13a):

- U6: > 150 cm of medium sand with climbing ripples
- U5: 110 cm of fine sand showing planar bedding and tiny ripples, with rare intercalations of thin indurated clay beds
- U4: 1 cm thick, indurated clay bed evolving laterally into a 10 cm thick mass-flow with sand boulders in a clay-silt matrix
- U3: 130 cm of silts with 5mm-thick, evenly-spaced laminae and thin, indurated, interbedded clay beds
- U2: 25 cm of coarse sand with climbing ripples
- U1: > 50 cm of fine sand with planar bedding

The thin indurated clay bed (U4, Fig. 7-13d) is particularly interesting, as it was deposited by decantation and marks thus a brief flow interruption in a water body, during which a small mass flow occurred, reworking previous silt and sand deposits (Fig. 7-13c). The thin clay bed of U4 is very continuous and can be followed on all the northwestern outcrops up to the southeastern part of outcrop 2, where it pinches out. U6 constitutes the second interesting feature and is constituted of more than 150 cm of climbing ripples. The high angle of climb (17°) indicates a high rate of net vertical deposition, related to a decelerating flow in a water body.

All our observations point out to subglacial meltwater circulations ending in a water body, which could be subglacial or periglacial. In the latter case, the flow-energy decrease between the gravel and sand formations (outcrop 1, Fig. 7-11) could indicate the switch from subglacial to proglacial glacio-lacustrine sedimentation during the retreat of the ice front.

Two km to the SW of Les Mossières, similar glacio-fluvial deposits are present at the Passoir gravel pit (519.0/152.5, commune of Saint-Livres). In 2002, a large N-S outcrop indicated poorly sorted gravel and sand layers with an arched, convex-upward bedding, suggesting an esker-like sedimentation. The DEM does not show any esker morphology around the pit, but the relief could have been levelled by ice erosion and/or infilling of depressions by glacial sediments. In 2006, a W-E outcrop displayed an inverse fault affecting stratified glacio-fluvial sediments dipping towards the west (Fig. 7-14). The fault probably formed during a lateglacial readvance of the Rhone glacier over frozen glaciofluvial material (permafrost).

Discussion and conclusion

Two different interpretations were proposed for the BGC. According to Arn (1984), the sediments accumulated in a basin limited to the south and the east by two arms of the Rhone glacier and to the west by the Jura slope. This interpretation, however, is not compatible with the ridge morphology: meltwater coming from the glacier front and limited distally by a gentle inverse slope would have generated a wedge-shaped, flat-topped accumulation thinning towards the Jura, where sediments would onlap the pre-existing till deposits. Holzmann (1999) proposed a slightly different interpretation: the BGC was formed in an ice-marginal lake bounded to the west by the readvancing Jura glaciers. The glacio-fluvial ridges and mounds to the west of the BGC would be glacio-fluvial sediments deposited by the Jura glacier during its retreat, after the formation of
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Les Mossières' gravel pit (Bière, VD).

NE (30°) SW (210°)

Fig. 7-11 Mossières gravel pit - outcrop 1. Flow is out of the page. See Fig. 7-10 for location and text for explanations.
Fig. 7.12 Mossieres gravel pit - outcrops 2,3,4; flow towards the NW; U4 = thin indurated clay bed. See Fig. 7-10 for location.
Fig. 7-13 Mossieres gravel pit - details. **a)** zoom on the western part of outcrop 3. Flow towards the right. Units discussed in the text. **b)** Gravel unit at the bottom-left of outcrop 4. Gravels are poorly sorted and poorly bedded, with stratifications dipping gently to the right. **c)** U4 - small clay-silt mass flow including boulders of stratified sands. **d)** U4 - thin indurated clay bed. **e,f)** Outcrop 5 - climbing ripples in flow-parallel and flow-transverse section. Note their through-shaped geometry. See Fig. 7-10 for location.
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The BGC. This interpretation is also questionable because the maximum extent of the readvancing Jura glaciers only reached the foothill of the Jura range, more than 2 km away from the gravel accumulation (Arn, 1984), and the LiDAR data do not show any evidence of an older, more important readvance.

A model for the BGC formation must take into account all the previously exposed geomorphological and sedimentological observations. The restricted extent of the BGC and its relatively steep borders suggests an ice-dammed accumulation. The Mossières gravel pit clearly points to a spatial and chronological transition from a subglacial environment to a proglacial, lacustrine environment. The small SW-NE oriented esker segments to the west of the BGC indicate an early subglacial stage. The connectivity of these small eskers to the BGC main N-S ridge suggests that these morphologies formed at the same time. The absence of till on the top of the gravels, the absence of glacio-tectonism and its uneven surface indicate that the BGC was not overridden by a younger glacial readvance. Therefore, it was formed between the end of the WGM (~ 18'000 BP) and the Montosset readvance (16’000-14’800 BP; Arn, 1984).

Similar elongated accumulations showing a transition from a subglacial/glaciofluvial to an ice-marginal/glaciolacustrine environment have been described in many deglaciated areas such as Ireland (Warren & Ashley, 1994; Glanville, 1997), Scandinavia (Punkari, 1997; Mäkinen, 2003) and North America (Barnett et al., 1998; Russell & Arnott, 2003). In the Oak Ridge Moraine, a 150 km long glaciofluvial accumulation in southern Ontario, Russell and Arnott (2003) interpreted deposits similar to those of the Mossières gravel pit as “upper-flow-regime hyperconcentrated-flood-flow deposits emplaced under a regime of rapid flow expansion and loss of transport capacity”. All these examples formed in an interlobate position during deglaciation, between two shrinking ice lobes and may present connection with eskers (e.g., Warren & Ashley, 1994). Thinner ice in interlobate zones caused convergence of supraglacial and subglacial drainage and concentration of meltwater (Punkari, 1997).

Given the above-mentioned similarity with interlobate deposits and the near absence of Jura lithologies in the BGC, the formation of the latter is interpreted as having taken place between the active Rhone glacier to the east and a large dead ice body abandoned by the Rhone glacier to the west. The BGC is located near the Rhone glacier division into its northeastern and southwestern ice tongues. Because of this location, specific conditions probably led to a complex ice flow and finally to a complex deglaciation sequence, thereby allowing the preservation of a large dead ice body to the west of the accumulation.

7.2.3 Joux Valley esker

The Joux Valley esker is located along the northwestern side of the valley, west of Le Brassus (503.2/158.2). The esker morphology consists of a 3300 m long ridge accurately imaged by the LiDAR DEM (Fig. 7-15). At the southwestern extremity, the esker starts as a low (~3 m), sinusoidal, 350 m long segment located approximately at the valley center. Further eastward, the esker becomes wider and makes a left turn to join the valley side. It stretches along the valley side and reaches its maximum size (10 m high, 300 m wide) in a zone exploited for gravels. The esker widening probably reflects a meltwater flow increase resulting from the inflow of englacial meltwater tributaries or crevasses. It confirms the northeast flow direction already indicated by current ripples in the Le Carré gravel pit (Fiore et al.,
Fig. 7-15 The Joux Valley esker (hillshaded view of the LiDAR DEM). Inset map shows the geomorphological interpretation. See Fig. 7-1 for location.

Fig. 7-16 Cross-section in the Joux Valley esker at the “Le Carré” gravel pit, May 2005. Flow out of the page. On the left, the sub-horizontal level is due to a step in the excavation front. See Fig. 7-15 for location.
Further eastwards, the esker becomes lower and its width decreases progressively.

The sedimentology and architecture of the esker central part were investigated by Fiore et al. (2002) at the Le Carré quarry, using outcrop observations and ground penetrating radar surveys. Since these observations made in 2000, the Le Carré quarry outcrop has shifted towards the SW following recent gravel exploitation (Fig. 7-16). It is similar to the 2000 outcrop but gives us some new information. The esker infill is asymmetric, the layers dipping towards the SE in most of the section. The dip is stronger than in the section observed in 2000 and the layers are convex-upward, whereas they were concave-upward in the former outcrop. This indicates a complex macroform made of backset beds evolving downflow from convex-upward to concave-upward in cross-sections transverse to flow. The conduit was filled from right (NW) to left (SE) and the unconformity between the stratification and the topography is probably depositional, rather than erosional.

The LiDAR data do not show other traces of subglacial meltwater circulation in the Upper Orbe valley. High-resolution seismic reflection investigation showed the presence of a deep stratified sequence at the centre of Lake Joux, interpreted as a subglacial delta (Bruder, 2003).

To summarize, the LiDAR DEM high resolution confirms the presence of an esker along the northwestern flank of the Joux Valley. Contrary to the previous interpretation of Fiore et al. (2002), the esker morphology is preserved and was not drumlinized after its formation.

### 7.3 Moraines

#### 7.3.1 Lausanne-Geneva Moraines

Between Lausanne and Geneva, the lakeside is constituted of Molasse bedrock covered by glacial and glacio-lacustrine deposits (Weidmann & Arn, 2005). Larges and continuous fronto-lateral moraines are visible between Lausanne and Gland; they area associated to similar moraines on the opposite shore, near Yvoire (FR).

Between Gland (VD) and Geneva, the Molasse reliefs is discernable in the topography by the presence of elongated, rectilinear, broad and low ridges oriented parallel to the Alpine tectonic trend: WSW-ENE in the north to NW-SE in the south (Figs. 7-17 and 7-18). These ridges are most probably sandstone beds, separated by less competent clay intervals more easily eroded by the sub-parallel ice flow.

Numerous low and smooth moraines are superimposed over this structural relief (Figs. 7-17 and 9-2). Some of them are indicated on the existing geological maps but their low relief and their intersection with the SW-NE Molasse relief led to an inaccurate and incomplete mapping. In figure 7-18, a new map of these moraines is proposed, based on the low-cut filtering of LiDAR data. Moraines are numerous between Gland and Bellevue (GE) and on the plateau of Jussy, northeast of Geneva (510.0 / 121.0). Individual moraines are 150-300 m wide, 5-10 m high and present a slightly meandering path. Present drainage network is partly guided by the moraines up-slope border. Overall, the moraines form a concentric pattern corresponding to the successive fronts of the receding Rhone glacier. The latter was centred on the lake valley but not confined to it.

The Rhone glacier recessional moraines were associated to decreasing lake levels which were higher than today, as shown by the overlying lacustrine sediments. Therefore, it is probable that their low relief results from a subaqueous formation, by pushing of water-saturated sediments with a low friction angle. Following the moraine formation, the relief was further smoothed by the sedimentation of overlying lacustrine material and finally by human activity, particularly agriculture.

#### 7.3.2 Montosset morainic ridges

In the Bière area, the DEM shows another type of landform superimposed on top of the hummocky ribbed
moraines: numerous continuous, narrow, sub-parallel ridges with an average SSW-NNE orientation (Fig. 7-19). Their shape and concentric disposition, typical of recessional moraines, indicate a formation through oscillations of the Rhone glacier during its retreat towards the ESE. The influence of the Jura ice front is not considered because the moraines are more distant from the Jura than other Jura ice cap moraines. Besides, the intersection of morainic ridges at point 518.9/153.2 indicates an ice front pushing towards the WNW and reworking older ridges (Fig. 7-19).

Among these parallel moraines, the most prominent was formed during the Montosset readvance (Arn, 1984). According to the stratigraphic study of Arn (1984), this readvance measured at least 350 m, maybe up to 3 km. This author attributes the readvance to the Dryas IB (16’000-14’800 BP) and proposes a correlation with the Laconnex stage (SW of Geneva). The moraine is particularly prominent around the locality of Montosset (519.0/153.6), where a portion of the ridge has been removed for gravel exploitation. The LiDAR reveals the large extent of the Montosset moraine, which can be followed over a distance of 15 km from the Jura foothills near L’Isles (521.2/163.2) until the top of the La Côte escarpment, north of Rolle (514.7/148.4), with an orientation passing from N-S to NE-SW (Fig. 7-18). On the westernmost part of the La Côte plateau (509.0/145.5), the Montosset moraine reappears, with a W-E orientation. The elevation of Montosset moraine is quite constant, oscillating between 668 and
708 m, with a mean value of 710 m. This indicates a low gradient of the Rhone glacier ice surface.

The LiDAR also reveals that the Montosset moraine is flanked by smaller parallel moraines (Fig. 7-19). These landforms were not recognized before and are absent in the geological maps, probably because of their low relief and the partial forest cover. Their number is variable: they are absent in the northern part of the Montosset moraine whereas up to nine crests are visible at the latitude of the Ballens gravels (520.4/155.0). In this zone, they are symmetrical in cross-section, 40-50 m wide, 2-5 m high and occur over a 1 km wide zone, on both sides of the Montosset moraine. The latter is difficult to distinguish from the nearby ridges, as if the Montosset ridge itself and the five other ridges visible further north on the other side of the Montosset ridge, a total of 11 Montosset readvances can be observed in the area, revealing the highly oscillating character of the ice retreat. This number is a minimum, because other morainic ridges were probably eroded or covered by fluvial deposits. The lateral variations in the number of crests is not related to the local topography and remains enigmatic.

To summarize, the Lidar data could display subtle topographic variations and evidence three generations of landforms in the Bièvre area: 1) hummocky ribbed moraine formed subglacially, probably at the same time as the nearby Molasse drumlins; 2) a subglacial to periglacial interlobate glaciofluvial complex in the Ballens area, formed at the confluence of the Rhone and Jura glaciers; 3) eleven sub-parallel recessional moraines corresponding to oscillations during the Montosset readvance.

Fig. 7-19  a) Bièvre area: hillshaded view blended with low-cut filtered elevation (Gaussian filter with a standard deviation of 1250 m). b) Geomorphological interpretation; the ridge intersection indicated with a star (*) reveals that the ice front pushed towards the WNW. See Fig. 7-1 for location.
Analysis of DEMs allowed a detailed mapping of rock-drumlins, drumlins and ribbed moraines in Switzerland and adjacent areas. Among the numerous topographic attributes derived from the DEMs, the low-cut filter developed on purpose was the most efficient because it highlighted the glacial landforms in order to map them accurately. This constitutes the first study of drumlins at the scale of Switzerland and the first report of ribbed moraines under the Alpine ice cap.

The type of subglacial bedforms depends on the substratum lithology. Rock drumlins occur over the hard Molasse bedrock, while drumlins and ribbed moraines are present over soft Quaternary deposits. The first were shaped by the successive Pleistocene glaciations, while the second formed during the last deglaciation and maybe during the WGM.

Eskers were recognized only in the western part of Switzerland covered by LiDAR data. The apparent absence in the rest of Switzerland may result from the insufficient resolution of the DHM25, from the burying of the eskers under pro-glacial sandur deposits during ice retreat or from their exploitation for sand and gravel. In Western Switzerland, the LiDAR high resolution revealed numerous recessional moraines formed by the fluctuations of the ice front.

The association of subglacial landforms built under the WGM Alpine ice cap (tunnel valleys, rock drumlins, drumlins, ribbed moraines and eskers) had been reported so far only from larger ice sheets (Laurentide, Fennoscandian, Irish). Its recognition points to a similar subglacial system for the Alpine ice cap.

Drumlins, ribbed moraines and eskers indicate warm-based subglacial conditions during the deglaciation and probably during the WGM. On the other hand, glacial readvances have produced inverse faults in coarse sediments, indicating a frozen, cold-based margin (Fig. 7-14). The warm-based conditions at the ice sheet centre results from the accumulation of geothermal heat below an important ice thickness, whereas the cold-based conditions at the margin results from the predominant influence of the cold atmosphere through a thin ice cover. This thermal model is in agreement with subglacial reconstitutions of larger ice sheets (e.g., Sollid & Sorbel, 1994; Kleman & Hättestrand, 1999; Benn & Clapperton, 2000; Marshall & Clark, 2002).

The Alpine foreland constitutes a very interesting area for the study of past subglacial processes. Despite the small dimensions of the Alpine ice cap, subglacial landforms have the same size as those described in North America or Fenno-Scandinavia, formed under larger ice-sheets.

Future research could take benefit of the high-resolution LiDAR DEM for remaining cantons to complete/refine our morphological analysis. This morphological information could be combined in a GIS with the existing lithological and stratigraphical information available from the literature and numerous boreholes. Because each glacier pertaining to the Alpine ice-cap constitutes a relatively independent sub-system, comparison of these sub-systems would permit to separate global processes from local ones.
Part II

Western Lake Geneva
Seismic Stratigraphy
Chapter 9

Introduction to Part II

9.1 Aim of the study

This second part presents the seismic stratigraphy of the Petit-Lac (western Lake Geneva), based on a large set of high-resolution seismic reflection profiles. In this area, glacio-lacustrine and lacustrine sediments have been intensively studied, resulting in an accurate reconstitution of the lake and climate evolution since the deglaciation (Girardclos, 2001; Baster, 2002; Girardclos et al., 2005). However, regarding the thick glacial sequence, no comprehensive seismic interpretation was proposed. Based on previously acquired and new seismic data, the main goal of this project is to resolve the complex geometry of the Quaternary glacial deposits and to propose an accurate and reliable model of the sedimentary infill, in order to shed new light onto the glacial processes related to the past glaciations and onto the dynamics of the past Rhone glacier.

Two particular aspects of this infill will retain our attention: esker and glacial readvances. Eskers in western Lake Geneva were first recognized by André Pugin. The new seismic coverage presented here allows the mapping of these eskers and a better understanding of their depositional conditions. Most of the Pleistocene eskers ended in a water body, while most of present-day eskers are ending subaerially (Banerjee & McDonald, 1975). Therefore, the study of water-ending fossil eskers by seismic reflections offers a unique opportunity to understand their geometry and sedimentology. Besides, eskers are potential aquifers, and mapping their extent and connectivity is useful for groundwater management and the understanding of subsurface pollution propagation.

Glacial readvances in the Petit-Lac area have formed fronto-lateral moraines visible onshore, but they mainly left their signature in the lacustrine sedimentary record, forming large push-moraines. Seismic reflection provides a unique means to obtain information about the internal geometry of these glacial morphologies and about their stratigraphical and therefore temporal relationship over a wide area. The sublacustrine record is better preserved than the terrestrial record because it has been sealed by glacio-lacustrine sediments right after the glacier retreat and has not been altered by post-glacial processes.

Fig. 9-1  a) Location map of Lake Geneva and extent of the Würmian Glacial Maximum (WGM). b) Bathymetry of Lake Geneva and geological setting.
9.2 Structure of the study

After an introduction to the geological setting of Western Lake Geneva and a review of the previous seismic investigations conducted there, chapter 10 exposes the seismic acquisition and processing methods. The seismic stratigraphy is presented in chapter 11 and these results are discussed in chapter 12. Finally, chapter 13 summarizes the conclusions and suggests some perspectives of research.

9.3 Geological setting

During Quaternary glaciations, the increasing erosion of the substratum by the ice streams led to the formation of numerous elongated perialpine lakes. During each deglaciation, these lakes acted as sediment traps. Each glacial cycle eroded most of the sediments of the previous one and further eroded the substratum. Today, perialpine lakes contain glacial and lacustrine sediments deposited during the last deglaciation and constitute therefore important geological archives to understand past glacial processes and climatic variations.

Lake Geneva (Le Léman in french) is the largest lake of Western Europe. It lies in the Alpine foreland between the Alps and the Jura range, on the course of the Rhone River (Fig. 9-1). Its northern shore and its two extremities are Swiss and the southern shore is French. The water level mean altitude is 372 m. The present study focuses on the western part of Lake Geneva (Fig. 9-2), known as “Petit-Lac” due to its small dimensions compared to the eastern part of Lake Geneva, known as “Grand-Lac” (Table 9-1). The Petit-Lac is elongated and narrow, with a maximum depth of 76 m and a length of 23 km, while the Grand-Lac consists of a deep and broad basin.

The Lake Geneva basin is cut into the Molasse bedrock, which consists of sandstones and marls eroded from the Alps and deposited in the northern Alpine foreland basin during the Oligocene and Miocene. Tectonically, the western part of Lake Geneva is located on the flat lying and little deformed plateau Molasse, but east of Lausanne, it crosses the subalpine front an cuts into the more internal subalpine Molasse characterized by closely imbricated Alpine thrust faults. During the Würmian maximum, the Rhone glacier covered the Geneva area with more than 800 m of ice (Jäckly, 1970; Monjuvent & Nicoud, 1988a). For more details about the Rhone glacier configuration and the timing of the glaciation, please refer to section 1.2.

As far as its glacial history is concerned, the Petit-Lac is a fjord-type lake resulting from the combined action of ice and water in a subglacial to glacial valley. During the periods of high ice level, the Petit-Lac bedrock valley acted as part of the subglacial meltwater drainage. During ice-retreat, when the Rhone glacier was narrower, ice flowed primarily through this valley in a fjord-like setting, even if the ice extended outside the valley, as shown by the spatial distribution and orientation of the lateral moraines (Fig. 9-2). Today, this valley is filled by a thick sequence of glacial, glacio-lacustrine and lacustrine deposits, mainly deposited during the last glacial cycle, as deposits from older glacial cycles have been almost completely eroded (Amberger, 1978; Wildi & Pugin, 1998).

The onshore Quaternary stratigraphy around the Petit-Lac is complex because of the glacial sediments’ heterogeneity, extreme local variability and the absence of simple stratigraphic rules. However, for the Geneva area, a general stratigraphic sequence has been established (Amberger, 1978; Maystre & Vergain, 1992). The Quaternary sequence lies directly upon the eroded Molasse bedrock, which is dissected by a network of deep and narrow tunnel valleys (Amberger, 1978; Signer, 1996). Between these valleys, the interfluves may be prominent and form large hills which are slightly elongated in the ice-flow direction. On top of the bedrock, the oldest unit is an overcompacted till (Moraine basale inférieure) dating from the Rissian or the early Würm (Reynaud, 1982; Maystre & Vergain, 1992). It is often overlain by a lignite-rich marl (Marnes à lignites) attributed to an early Würm interstadial. These two former units have been heavily eroded during the Würm glaciation and are very discontinuous.

On top of this erosion surface lies the Alluvion Ancienne, a thick and extended unit of gravel and sands, covering the whole Geneva area except the high-relief Molasse hills. It corresponds to a proglacial outwash plain (sandur) deposited in front of the advancing Rhone glacier at the beginning of the Würm glaciation or at the beginning of the most important Würm stadium, older than 38400 BP (Monjuvent & Nicoud, 1988b). This unit probably encompasses distinct gravel units (M. Meyer, pers. comm.). The subsequent glacial activity has locally eroded the Alluvion Ancienne, especially in the prolongation of the Petit-Lac valley. It is overlain by a complex and discontinuous succession of till and glaciolacustrine units deposited during the Würm deglaciation.

<table>
<thead>
<tr>
<th>Lake basin</th>
<th>Surface (km² / %)</th>
<th>Volume (km³ / %)</th>
<th>Maximum depth (m)</th>
<th>Length (km)</th>
<th>Maximum width (km)</th>
<th>Catchment area (km² / %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Petit-Lac</td>
<td>81.2 / 14</td>
<td>3 / 4</td>
<td>76</td>
<td>23.3</td>
<td>4.7</td>
<td>325 / 4.4</td>
</tr>
<tr>
<td>Grand-Lac</td>
<td>498.9 / 86</td>
<td>86 / 96</td>
<td>310</td>
<td>49</td>
<td>13.8</td>
<td>7070 / 95.6</td>
</tr>
<tr>
<td>Lake Geneva (total)</td>
<td>580.1 / 100</td>
<td>89 / 100</td>
<td>310</td>
<td>72.3</td>
<td>13.8</td>
<td>7395 / 100</td>
</tr>
</tbody>
</table>

Until now, four deglaciation stages have been proposed in the Geneva area: Laconnex, Geneva, Coppet, Nyon and Thonon (Fig. 9-1) (Moscariello et al. *cum biblio*, 1998; Chapron, 1999). This retreat was associated with successive decreasing lake levels ranging from 470 m asl to the 372 m present-day level, leading to the formation of several generations of terraces (Burri, 1981; Moscariello et al., 1998). As a result, most of the Western Lake Geneva area is now covered with clayey lacustrine sediments, smoothing heavily the morphology inherited from the glaciations.

The two first retreat stages (Laconnex and Geneva) are associated with proglacial deltaic deposits (Reynaud, 1982; Moscariello et al., 1998). Based on geomorphological arguments, they have been tentatively associated respectively with the 470 m and 405 m levels, with an intermediate lake level at 430 m between the two stages (Moscariello et al., 1998).

The third and fourth main stages (Coppet and Nyon) are marked onshore by the presence of fronto-lateral moraines (Fig. 7-18). In the lake, they are associated with acoustically transparent sediment wedges visible on longitudinal lacustrine seismic reflection profiles interpreted as till-tongues (Moscariello et al., 1998), following the model developed by King et al. (1991). The last stage (Thonon) was inferred from a till-tongue recognized by Chapron (1999) on a sparker seismic section.

For the Geneva basin, Monjuvent proposed a finer division of the deglaciation into six steps labeled ‘step 3’ to ‘step 8’, based mainly on the presence of large proglacial deltas (Donzeau et al., 1997). Each step is characterized by the position of the retreating Rhone glacier and one to three associated proglacial lakes. The successive lakes show decreasing levels ranging from 650 m (beginning of step 3) to 410 m (step 8). During each step, tills were deposited below the glacier while glacio-lacustrine sediments were accumulating in peri-glacial lakes and associated deltas.

Datations of glacial sediments in the Geneva bay indicate that ice was sill present at 18940±210 yrs 14C BP (Moscariello, 1996) whereas in the Lausanne area two datations of lacustrine terraces indicate that the lake was present at least since 13 kyr 14C BP (Arn, 1984; Gabus et al., 1987). These datations indicate that the deglaciation of Lake Geneva occurred between 19 and 13 kyr 14C BP (Girardclos, 2001).

### 9.4 Previous seismic studies

High-resolution offshore seismic reflection is a powerful tool for the investigation of Pleistocene glacial deposits. In the last decades, numerous studies have been conducted worldwide on continental margins, fjords or lakes, with various seismic techniques ranging from small single-channel 2D surveys to large multi-channel 3D surveys (Davies et al., 1997).

Since the 70’s, numerous seismic studies have been conducted on perialpine lakes. Finck et al. (1984) investigated the amount of bedrock overdeepening and the seismic velocities of 17 perialpine lakes, through seismic reflection and refraction. A seismic survey was conducted on Lake Zürich (Giovanoli et al., 1984), which is of particular interest because of its association with a borehole crossing the entire Quaternary sequence (Hsu et al., 1984; Lister, 1984b). In lakes Annecy and Le Bourget (France), the seismic study of Van Rensbergen (1996) led the way to a series of papers on their glacial and post-glacial deposits (Van Rensbergen et al., 1998; Chapron, 1999; van Rensbergen et al., 1999; Beck et al., 2001).

Regarding Lake Geneva, the first seismic survey was shot by Houbolt and Jonker (1968) and focused on the Rhone delta sediments, at the eastern extremity of Lake Geneva. The next survey was conducted with a boomer source on Lake Geneva western part (Vernet & Horn, 1971) and was later extended to the entire lake (Vernet et al., 1974), resulting in a series of isopach maps of the main seismic stratigraphic units. However, the relatively low penetration of the boomer source made the interpretation of the lowest units difficult. This led to the incorrect
indentification of faults in the Molasse bedrock, which were not recognized by subsequent studies (e.g., Moscariello et al., 1998; Chapron, 1999).

Several research projects at the University of Geneva (Forel Institute and Department of Geology) led to the acquisition of more than 900 km of seismic lines on Lake Geneva with five different sources ranging from high to very high resolution: Airgun 5 and 1-inch, Sparker, Impactor and Echosounder. Sources of lower resolution were used to study the Molasse bedrock (e.g., Morend et al., 2002) or the complete Quaternary sequence (e.g., Moscariello et al., 1998), whereas those of higher resolution were used to study the glacio-lacustrine and lacustrine sediments (e.g., Girardclos, 2001; Baster et al., 2003). Information concerning these surveys is summarized in Table 10-2.

Moscariello et al. (1998) studied the glacial history of the Petit-Lac southwestern termination near Geneva. These authors defined 5 lithostratigraphic/sismostratigraphic units based on the Geneva bay cores. Through a set of 5-inch seismic profiles, the lithostratigraphy was extended to the entire Petit-Lac. However, seismic data in the core area is of poor quality, due to gas blanking and numerous multiples caused by the shallow water depth. Also, longitudinal seismic lines show important lateral variations of the seismic units. Therefore, the correlation of these cores with the seismic record is uncertain.

Based on a small set of sparker lines, Chapron (1999) also investigated the sedimentary structure of Western Lake Geneva and proposed a general seismic interpretation. In the Grand-Lac, several 2D and 3D surveys have been conducted by the Institute of Geophysics of the University of Lausanne, in order to study the Molasse bedrock and, to a lesser extent, the Quaternary deposits (Scheidhauer, 2003; Dupuy, 2005).
10.1 Existing seismic data

Seismic data used for this study were acquired with six different seismic sources (15-, 5- and 1-inch\(^3\) airgun, sparker, impactor and ecosounder), characterized by different frequency contents (Table 10-1, Fig. 10-1). These different sources are complementary ways to image the sediments: lower frequency sources permit to reach the deepest deposits, whereas higher frequency sources offer a better resolution for the shallower deposits. The airgun source energy is proportional to the cubic square root of the airgun volume. The 15- and 5-inch\(^3\) airgun pulses are therefore respectively 2.47 and 1.71 times more powerful than the 1-inch\(^3\), improving significantly their penetration in the subsurface.

Part of the seismic data used in this study was shot during previous campaigns conducted by members of the Forel Institute and Department of Geology of the University of Geneva, most of them under the supervision of André Pugin (Table 10-2, Fig. 10-2). A 15-inch\(^3\) airgun line was also kindly provided by the Institute of Geophysics of the University of Lausanne, allowing the connection with the seismic interpretation of Western Lake Geneva (Dupuy, 2005). The seismic profiles and the mapping of the seismic units are regrouped in appendices D and E, and the location of the seismic lines presented in this study is indicated on figure D-1.

All these data were loaded into a single seismic interpretation project in the Kingdom Suite software. Interpretation was based primarily on the dense grid of 1-inch\(^3\) and 5-inch\(^3\) airgun profiles. Sparker data are of excellent quality and provided useful images for the upper and intermediate sediments, but they could not be used for the generation of isopach/isochron contour maps because their positioning is very inaccurate, with up to 200 m of error. Due to their low seismic penetration, Echosounder and Impactor data were used only for the upper part of the glacial sediments.

### Table 10-1

<table>
<thead>
<tr>
<th>Source</th>
<th>Bubble canceling</th>
<th>Airgun 15/15-inch(^3)</th>
<th>Airgun 5-inch(^3)</th>
<th>Airgun 1-inch(^3)</th>
<th>Sparker</th>
<th>Echosounder</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of channels</td>
<td>66</td>
<td>11</td>
<td>11</td>
<td>1</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Channel spacing (m)</td>
<td>2.5</td>
<td>7</td>
<td>7</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Minimum offset (m)</td>
<td>7.5</td>
<td>7</td>
<td>7</td>
<td>8</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Shot spacing (m)</td>
<td>10</td>
<td>7 m</td>
<td>7 m</td>
<td>1.5 s (~ 2.3 m)</td>
<td>0.25 s (~ 0.38 m)</td>
<td></td>
</tr>
<tr>
<td>Nominal CMP fold</td>
<td>9</td>
<td>5 to 6</td>
<td>5 to 6</td>
<td>1</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Trace spacing (m)</td>
<td>1.25</td>
<td>3.5</td>
<td>3.5</td>
<td>~ 2.3</td>
<td>~ 0.38</td>
<td>~ 0.38</td>
</tr>
<tr>
<td>Frequency (Hz)</td>
<td>50-330-650</td>
<td>50-150-900</td>
<td>120-400-1600</td>
<td>150-1000-1800</td>
<td>1500-1750-2000</td>
<td></td>
</tr>
<tr>
<td>min-dominant-max</td>
<td>Theoretical vertical resolution (m) for a velocity of 1600 m/s</td>
<td>1.2</td>
<td>2.66</td>
<td>1</td>
<td>0.4</td>
<td>0.23</td>
</tr>
<tr>
<td>Theoretical horizontal resolution (m) for depths of 50 and 300 m.</td>
<td>20.0-49.0</td>
<td>23.0-56.6</td>
<td>14.1-34.6</td>
<td>8.9-21.9</td>
<td>6.8-16.6</td>
<td></td>
</tr>
</tbody>
</table>

*Table 10-1* Acquisition parameters and seismic resolution of the seismic sources used in this study. Horizontal and vertical resolution calculations are based respectively on Ralleigh's criterion \(\lambda/4\) and on the first Fresnel zone radius \((\lambda * depth/2)\)^{0.5} (Yilmaz, 1987).
Chapter 10 - Seismic acquisition and processing

Fig. 10-1 Amplitude spectra of the different seismic data used in this study. For the 1-inch³ airgun, note the variations in amplitude spectrum between surveys. FL stands for “frequency loss” and represents a range of frequencies cancelled by the air/water ghost reflection. The attenuated frequency can be calculated with the following formula: 0.5 * water_velocity / airgun_depth. For an airgun depth of 0.8 m, this formula gives a frequency of 900 Hz corresponding to the frequency spectrum shape.

10.2 Acquisition

More than 200 km of seismic data were acquired specifically for this study between 2002 and 2005, with a 1-inch³ airgun source and a 11-channels streamer (Figs. 10-3 and 10-4, Table 10-2). Seismic data was recorded with a Geometrics R-48 seismograph, except for the last survey (AG1_JF_05), for which a Geometrics Geode connected to a laptop PC was used. The acquisition was conducted either from the Forel Institute research vessel “La Licorne” or from the smaller “Lehman” boat. We used a Global Positionning System (GPS) device to navigate, calculate the shot interval and record the shot coordinates. Additional acquisition parameters are summarized in Table 10-3.

To allow a 3D reconstruction of the subsurface architecture, lines were acquired in directions transverse and parallel to the lake axis. Adding the newly acquired data to the previously acquired data (AG1_AP_00 and AG1_MB_02 in Table 10-2), the 1-inch³ airgun dataset for the study area consists of 78 high-resolution seismic lines totaling a distance of 240 km.

A subset of 37 lines (77 km) in the Coppet area constitutes a pseudo-3D survey (also called 2.5D survey), i.e., a set of parallel and evenly-spaced 2D seismic lines which, when interpreted and interpolated, provide a 3D model of the subsurface. This pseudo-3D was shot perpendicularly to the main geological structures (i.e., perpendicularly to the lake axis) with an average line spacing of 126 m over an area of 9.5 km².

Acquisition parameters for other datasets are summarized in Table 10-1. More details can be found in the publications listed in Table 10-2 for the datasets acquired by the UNIGE/IAF, or in Dupuy (2005) for the 15-inch³ airgun data.

10.3 Data quality and artifacts

Data quality

The data quality can be affected by several parameters and varies between surveys. The lake condition during acquisition is one of the main problems. Waves slap against hydrophones and buoys, generating random noise. This random noise cannot be properly removed because it is not coherent nor restricted to a specific frequency band. It is therefore important to check the streamer before acquisition to make sure that the buoys do not hit the hydrophones. Longitudinal profiles acquired in the same direction as the Rhone River through Lake Geneva are usually less noisy because of the smaller water friction on the system. In the case of very noisy channels, those had to be suppressed.

The quality of the seismic recorder can also affect the data. We noted that the data recorded with the Strataview seismograph were noisier and had a narrower frequency spectrum than the data recorded with the more recent Geode seismograph (survey AG1_JF_05). This difference is reflected in their frequency spectrum (Fig. 10-1): the Geode spectrum is smoother and broader than the Strataview spectrum.

Multiples

Reverberation between the lake bottom and the lake surface generates multiples. As the strength and frequency of multiples are more important in shallow water near the lake borders, we avoided generally the seismic acquisition in water depth smaller than 30 m (0.042 seconds TWT). During interpretation, the position of the first and second multiples was calculated by multiplying respectively by two and three the two-way time of the lake bottom horizon, in order to avoid a confusion of the multiples with primary reflections.
**Sideswipe events**

Because of the rapid lateral variations in the glacial sediments, longitudinal lines are necessary to link transversal lines. As they are not shot perpendicularly to the geological structures, they are more prone to be affected by sideswipe events, i.e., reflections produced by features out of the plane of the seismic section. During interpretation, these sideswipe events may be misleading.

Minor time shifts (< 3 ms) were also observed during interpretation at the crossing points between surveys. These time shifts result (1) from tiny differences in offset and in source/hydrophone depths and (2) from the fact that 2D migration is based only on the dip components in the in-line direction, leading to a bad time tie of dipping reflections at line intersections (Sheriff & Geldart, 1995).

**Gas blanking**

In part of the study area, seismic records are affected by gas blanking (Figs. D-2 and E-1), attributed to methane produced by anaerobic bacteria in the uppermost and youngest lacustrine sediments. As gas bubbles are compressible and behave elastically, they continue to resonate and emit acoustic energy long after their initial compression by the source wavefront, producing incoherent noise beneath the gas-rich horizon on the seismic record (Davies et al., 1997). Depending on gas concentration, the record may be slightly to heavily degraded, hiding in this latter case all reflections below the gas-rich horizon. However, in most cases, the deeper part of the record is still interpretable because gas blanking decreases downward. Gas blanking affects principally the southernmost part of the Petit Lac.

**10.4 Airgun data processing**

Data acquired with the 1-inch\(^1\) and 5-inch\(^1\) airguns were processed on PC with Seismic Unix (SU) and shell scripting (Stockwell & Cohen, 2002). SU is an open source package providing a complete and powerful seismic processing solution. A guide to the installation and basic use of SU under Windows is available in Appendix B.

The processing flow (Table 10-4) is based on that developed by A. Pugin between 1996 and 2000 at the University of Geneva. All the processing steps originally performed with various softwares (Eavesdropper, Vista and home-made softwares) have been ‘translated’ into SU commands. The use of a single processing package avoids tedious and time-consuming conversion between different software-specific seismic formats. It offers more flexibility and permits to automatize operations for faster processing. Furthermore, some processing steps have been added to the original processing flow of A. Pugin: velocity analysis, spherical divergence compensation and diversity stacking. All the airgun data were processed with this new processing flow. The practical details are presented in Appendix C.

**Geometry assignment**

The geometry assignment consists in writing the offset and common depth-point (CDP) number in the header of each trace. The offset is the distance between the source and the hydrophone and is channel-dependent. The CDP number indicates the theoretical reflection point for a given source-receiver pair and is located at half-distance between the source and the receiver. In the next processing steps, the offset will serve to calculate the normal move out (NMO) and the CDP number will serve to gather the traces to be stacked.

**Trigger delay**

For each shot, the airgun explodes 10 ms after the seismograph starts to record. In order to correct this trigger delay, the first 10 ms of each trace must be removed.

**Spherical divergence compensation**

Absorption of the seismic energy by the earth and the spherical divergence of the source energy lead to a loss of amplitude with depth (Claerbout, 1976). To correct this, a data independent scaling function was applied by multiplying each sample amplitude by its time in seconds. This gaining function proved to be well adapted to our data. Thanks to this processing step, the amplitude of the reflections becomes depth-independent, allowing an unbiased interpretation of the seismic facies. On noisy data, the signal decreases with depth, but not the random noise. Thus, the gain enhances that noise in the deepest part of the profiles. No automatic gain control (AGC) was applied to preserve the relative amplitude of the reflections.

**First arrival mute**

The first arrival mute removes the first milliseconds of the traces affected by the wave travelling directly from the source to the hydrophones. Some lines were not muted to preserve the reflections in the shallow areas near the lake border. In that case, the direct wave is attenuated anyway by the gain and the diversity stack, which gives less weight to the anomalous high amplitudes.

**Frequency filtering**

The bandpass frequency filtering suppresses the undesired frequencies, especially the very low frequencies. It is defined by four values (f\(_1\) to f\(_4\)) which form a trapezoid on the frequency spectrum. The ratios f\(_2\)/f\(_1\) and f\(_4\)/f\(_3\) must be larger than √2 to prevent artifacts. The following
Fig. 10-2 Maps of the seismic lines used in this study.
## Chapter 10 - Seismic acquisition and processing

### Table 10-2

List of seismic surveys acquired on Lake Geneva since 1996 by the University of Geneva (Forel Institute and Department of Geology). Surveys sorted by seismic source, from lowest to highest frequency, and by chronology. AWI = Alfred Wegener Institut; LGCA = Laboratoire de Géodynamique des Chaînes Alpines; RCMG = Renard Centre of Marine Geology of Gand.

<table>
<thead>
<tr>
<th>Source</th>
<th>New survey name</th>
<th>Informal survey name</th>
<th>Goal and Location</th>
<th>Acquisition Date</th>
<th>Number of lines or Grid dimension</th>
<th>Distance acquired (km)</th>
<th>Acquisition leader(s)</th>
<th>Publications</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5 in.³ Airgun</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AG1_AP_97</td>
<td>Grand Lac 1 Airgun</td>
<td>Transverse lines across the Grand Lac</td>
<td>3 + 6.3.1997</td>
<td>4</td>
<td>36.2</td>
<td>A. Pugin</td>
<td>Girardclos (2001)</td>
<td>●</td>
</tr>
<tr>
<td>AG1_AP_00</td>
<td>CoppetDD</td>
<td>Pseudo-3D survey in the Coppet area to image eskers. 14 lines numbered 1920 to 4546</td>
<td>14.03.2000</td>
<td>14</td>
<td>31.4</td>
<td>A. Pugin</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
</tr>
<tr>
<td>AG1_MB_02</td>
<td>Milan AWI</td>
<td>Test to compare the Forel/UNIGE multi-channel acquisition system with AWI single-channel system. 30.05.2002</td>
<td>1</td>
<td>5.9</td>
<td>M. Beres, AWI</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
<td></td>
</tr>
<tr>
<td>AG1_JF_02</td>
<td>CoppetJ</td>
<td>Extension of the Pseudo-3D (AG1_AP_00) toward the south. 26-27.06.2002</td>
<td>23</td>
<td>46.1</td>
<td>J. Fiore, M. Beres</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
<td></td>
</tr>
<tr>
<td>AG1_JF_03</td>
<td>Aout03</td>
<td>Longitudinal lines in the Petit Lac to link the existing transverse lines and image the longitudinal section of elongated sedimentary bodies. 05-07.08.2003</td>
<td>14</td>
<td>74.6</td>
<td>J. Fiore, M. Beres</td>
<td>Girardclos et al. (2005)</td>
<td>● ● ●</td>
<td></td>
</tr>
<tr>
<td>AG1_JF_04</td>
<td>J04</td>
<td>Complete the coverage of the Petit Lac. 05-06.04.2004</td>
<td>9</td>
<td>23.8</td>
<td>J. Fiore, M. Beres</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
<td></td>
</tr>
<tr>
<td>AG1_JF_05</td>
<td>JF05</td>
<td>Complete the coverage of the Petit Lac. 25-27.5.2005</td>
<td>17</td>
<td>57.3</td>
<td>J. Fiore, M. Beres</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
<td></td>
</tr>
<tr>
<td>Bathy-1000 Echosounder (ODEC)</td>
<td>IMP_SG_98</td>
<td>Girardclos</td>
<td>Shot over pre-existing Echosounder lines (es.sg.97) to compare the sources</td>
<td>autumn 1998</td>
<td>6</td>
<td>8.2</td>
<td>S. Girardclos, A. Pugin</td>
<td>Girardclos (2001)</td>
</tr>
<tr>
<td>IMP_IB_99</td>
<td>Baster Impactor</td>
<td>Shot over pre-existing Echosounder lines (es.ib.97) to compare the sources</td>
<td>February 1999</td>
<td>3</td>
<td>10.9</td>
<td>I. Baster, A. Pugin</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
</tr>
<tr>
<td>ES_SG_97</td>
<td>Girardclos Echosounder</td>
<td>Holocene seismo-stratigraphy of the Hauts-Monts area</td>
<td>summer 1997</td>
<td>4 x 3.5 km grid, 260 m line spacing + 1 x 1 km grid, 65 m line spacing</td>
<td>106.0</td>
<td>S. Girardclos, A. Pugin</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
</tr>
<tr>
<td>ES_IB_98</td>
<td>Baster Echosounder</td>
<td>Holocene seismo-stratigraphy of the Promenthouse delta</td>
<td>summer 1997 + spring 1998</td>
<td>Grid with a 300 m line spacing</td>
<td>110.0</td>
<td>I. Baster, A. Pugin</td>
<td>Girardclos (2001)</td>
<td>● ● ●</td>
</tr>
</tbody>
</table>
trapezoid frequencies were used: 120/180-1200/1800 for 1-inch³ data and 80/120-1200/1800 for 5-inch³ data.

**Predictive deconvolution**

The air bubble released by the airgun expands until the water pressure compensates the air pressure. At this time, the bubble collapses. When the air pressure gets higher than the water pressure a second expansion takes place. The balance between the bubble air pressure and water pressure yields thus to a serie of oscillations decreasing in amplitude. On the seismic data, the bubble collapse appears as a second, weaker pulse known as bubble effect or bubble ghost. It occurs after a delay of around 25 ms for the 1-inch³ and 30 ms for the 5-inch³ airgun. Its amplitude is higher when the airgun is fired at a higher pressure. As the bubble effect delay is constant for a given acquisition system, it can be easily predicted and removed (or at least strongly attenuated) by predictive deconvolution, also known as statistical deconvolution or Wiener predictive error filtering (Benz, 1999). The deconvolution must be applied after frequency filtering and before the NMO correction, as this latter step distorts the wavelets. To estimate the appropriate deconvolution parameters, and especially the bubble effect delay, an autocorrelation analysis must be done previously.

**Velocity Analysis and NMO**

Velocity analysis yields the stacking velocities necessary to NMO correction, or flattening of the CDP gathers. It is especially important for high-frequency data (e.g., 1-inch³), which are more sensible to the NMO correction than lower frequency data. As the signal wavelength is short, individual traces have to be aligned precisely, otherwise reflections get out-of-phase and cannot be stacked properly. Ideal CDPs for velocity analysis should present numerous sub-horizontal reflections located at various depths. Because these conditions are not encountered on every point along a given profile, the CDPs to analyse were first selected on a preliminary section stacked with a constant velocity of 1500 m/s. A semblance analysis for a velocity range was subsequently performed for the selected CDPs. The velocity analysis presents three advantages over the constant velocity stack: the stack quality is good for both shallow and deep reflections, information about the sediment velocity is obtained and multiples are attenuated.

Lower frequency data (e.g., 5-inch³) have a longer wavelength and tolerate less accurate stacking velocities. Therefore, we used two constant velocity stacks for the 5-inch³ data: 1435 m/s and 1700 m/s. The first velocity of 1435 m/s corresponds to the water velocity (Coppens, 1981) and is satisfying for the upper part of the Quaternary infill. The second velocity of 1700 m/s stacks correctly the lowermost part of the Quaternary infill and the bedrock. In *Kingdom Suite*, both stacks were loaded and switched during interpretation.

**Stacking**

Stacking refers to the addition of all the traces pertaining to the same CDP to obtain a zero-offset section. It increases significantly the signal/noise ratio and attenuates multiples. The signal to noise ratio improvement is proportional to the square root of the fold (i.e., the number of traces in the gather). Our streamer has 11 channels, an unusual uneven number resulting from a fabrication error. As a consequence, the fold alternates between 5 and 6 on the stacked sections. Two stacking methods were used: normal stack (SUSTACK) and diversity stacking (SUDIVSTACK). Diversity stacking is a noise reduction technique. Prior to stacking, each trace is scaled by the inverse of its average power over a user-defined time window. The composite trace is then renormalized by dividing by the sum of the scalers used. This stacking method is very effective for data with an overall good signal/noise ratio affected randomly by high-amplitude noise (spikes).

**Migration**

Phase-shift migration is an essential step which collapses diffractions (convex-upward hyperbolae) and replaces seismic events at their correct position. A constant migration velocity of 1500 m/s was used. Spikes present in the original data produce large convex-downward hyperbolas in the migrated section, called “smiles”.

**3D cube generation**

The 37 lines of the pseudo-3D seismic survey have been regrouped into a single 3D cube. Compared to a set of 2D lines, the 3D cube presents the following advantages for seismic interpretation: easy navigation
Chapter 10 - Seismic acquisition and processing

Acquisition geometry and numbering of station, channel and CMP for the 1st and 2nd shot of an airgun seismic line. The first 11 traces have virtual coordinates, expressed as negative numbers. Note that the CMP number is equal to shot-station + receiver-station, and is increased by 2 between two successive shots.

Fig. 10-3 Acquisition geometry and numbering of station, channel and CMP for the 1st and 2nd shot of an airgun seismic line. The first 11 traces have virtual coordinates, expressed as negative numbers. Note that the CMP number is equal to shot-station + receiver-station, and is increased by 2 between two successive shots.

Fig. 10-4 Typical airgun seismic acquisition system on Lake Geneva with the “Lehman” boat.
between inlines, visualisation of time-slices, crosslines and oblique lines (Fig. 10-5), creation of movies scanning the entire cube in any direction and direct generation of a 3D surface from the picked horizons (no need to create a grid). Given the regular spacing of the original 2D lines, the coordinates of the cube inlines fit very well the original coordinates, as illustrated in figure 10-6. The practical details for the 3D cube construction are explained in Appendix C.6.

**Coordinates processing**

During processing, coordinates of the shot points were converted from the latitude/longitude system WGS84 to the Swiss grid CH1903 with the Excel spreadsheet wgs84.xls (Appendix F). It is important to note that the first GPS coordinate does not correspond to the first trace recorded but to the first source location. As illustrated by figure 10-3 for our 11 channels system, the first coordinate corresponds to CMP 224, while CMP numbering starts at number 213. CMPs 213 to 224 do not have coordinates. Added to these missing coordinates, the first 2 or 3 coordinates of each lines had to be suppressed because they were wrong, probably because of old coordinates remaining in the GPS buffer. For a correct shot-point/coordinate relationship in Kingdom Suite, these missing coordinates were taken into account when defining the “starting shotpoint”. Based on this information, Kingdom Suite extrapolates automatically the missing coordinates.

**10.5 Sparker data processing**

Sparker data were already processed by A. Pugin, but three post-stack processing steps have been added: a low-cut filter to remove vertical bands, a linear gain to compensate for spherical divergence and phase-shift migration. For the migration, the shot interval is mandatory but it is not constant for sparker data, as the shot interval is function of time, not of distance, and depends therefore on the boat speed. Our tests have shown that a shot interval of 2.3 m produced the best results. Processing of the sparker coordinates is exposed in appendix C.7.

**10.6 Gridding**

For the isochron and isopach maps, seismic data were first interpolated in SMT Kingdom Suite by triangulation with linear interpolation. This algorithm consists in a simple triangulation without extrapolation. It is faithful to the original data and particularly efficient on data sets containing dense data in some areas and sparse data in

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**Table 10-4** Airgun 1-inch¹ processing flow.

<table>
<thead>
<tr>
<th>Steps</th>
</tr>
</thead>
<tbody>
<tr>
<td>SEG2 to SEGY conversion</td>
</tr>
<tr>
<td>SEGY to SU conversion</td>
</tr>
<tr>
<td>Geometry assignment</td>
</tr>
<tr>
<td>Trigger delay correction</td>
</tr>
<tr>
<td>First arrival muting</td>
</tr>
<tr>
<td>Bad channels removal</td>
</tr>
<tr>
<td>Spherical divergence correction (gain)</td>
</tr>
<tr>
<td>Predictive deconvolution</td>
</tr>
<tr>
<td>Bandpass filtering</td>
</tr>
<tr>
<td>CMP sorting</td>
</tr>
<tr>
<td>NMO correction</td>
</tr>
<tr>
<td>Stacking or Diversity stacking</td>
</tr>
<tr>
<td>Phase-shift time migration</td>
</tr>
<tr>
<td>Pseudo-3D cube construction</td>
</tr>
<tr>
<td>SU to SEGY conversion</td>
</tr>
</tbody>
</table>

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¹¹-inch = 2.5 cm
other, like 2D surveys. Its application is very simple as the only required parameter is the cell size of the output grid.

The interpolation process is always a compromise between honouring the data and obtaining a smooth and pleasant surface. Larger grid cells produce smoother surfaces but do not honour correctly the rapid lateral variations of the data. We used a grid cell size of 50 m by 50 m to allow a compromise between the dense sampling on the 2D seismic lines (3.5 m) and the large distance between 2D lines (from 125 m in the pseudo-3D area up to 1500 m in the northern part of the Petit Lac).

After the interpolation, we applied two successive passes through a 5x5 median filter to remove or attenuate aberrant values due to inaccurate line crossing or picking errors. For each cell, median filtering calculates the median value of all pixels in the selected neighborhood, a 5x5 square in our case. The median filter has the advantage to remove impulse noise and to preserve realistic values when the filter straddles a steep slope (Russ, 1999).

Fig. 10-6 Comparison of the pseudo-3D cube location with the original 2D lines.
Chapter 11

Seismic Stratigraphy

11.1 Introduction

Seismic interpretation was conducted on PC with the SMT Kingdom Suite software, based on both migrated and unmigrated seismic sections, the latter preserving diffractions. Auto-tracking of seismic-horizons was used for the lake bottom and the lacustrine reflections. Deeper horizons were picked manually because of the poor lateral continuity of the reflections.

Based on strong reflections, unconformities or sudden changes in seismic facies, the sedimentary sequence of Western Lake Geneva was subdivided into 14 seismic units, each one representing a volume of sediments with similar acoustic properties deposited in the same time interval. For the digital interpretation, a horizon was assigned to the top of each seismic unit. In the case of tangential joining of erosive surfaces, we have followed the upper erosive surface and truncated the lower one.

Further to the subdivision into seismic units, special attention was paid to the variations of seismic facies. Seismic facies can be divided in two main categories: chaotic and stratified. The first category is common in the glacial environment, where till or poorly organized ice-marginal deposits lead to a “chaotic” or “semi-transparent” facies. In this type of facies, diffraction hyperbolas visible on unmigrated seismic sections are particularly important as they indicate boulders in the sediments. The interpretation of such facies is often ambiguous, because they can originate from different glacial processes and sediments (Davies et al., 1997; Syvitski et al., 1997).

The second category of seismic facies concerns deposits presenting some internal stratifications. In these facies, the interpreter’s attention focuses on the reflection’s amplitude, lateral continuity and spatial frequency and on their arrangement (e.g., parallel, undulatory, cross-cutting). Seismic facies analysis is intimately related to the subdivision into seismic units because some of the seismic units are differentiated on their seismic facies.

The absence of deep cores in the investigated area does not permit a direct relation between geology and seismic stratigraphy. Therefore, interpretation of the seismic unit facies and geometry was based on the comparison with the sedimentary sequence of other perialpine lakes – Lake Zürich (Lister, 1984b), Lake Annecy (Van Rensbergen et al., 1998) – and on the current knowledge of glacial processes. Seismic studies in fjords and in polar shelf regions present many similarities with fjord-lakes and can be used as reference, taking into account the differences between the marine and lacustrine environment.

The geometry and acoustic properties of each seismic unit are presented below from bottom to top, illustrated by a selection of key seismic profiles (Appendix D) and by isochron and isopach contour maps (Appendix E). Isochron maps indicate the two-way travel time in milliseconds (TWT) between the lake surface and a given stratigraphic level, while isopach maps indicate the TWT between two stratigraphic levels. If these maps were converted from time to vertical distance, they would indicate respectively the depth and the thickness of the seismic units.

Based on the geometry and acoustic properties of each unit, the lithology and depositional setting of each unit is

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**Fig. 11-1** Ambiguous stratigraphic relationship between reflections: the absence of stratifications within glacial units makes difficult to choose between an erosive and a depositional interpretation.
<table>
<thead>
<tr>
<th>Unit</th>
<th>Seismic Facies</th>
<th>Facies description</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>U12-U14</td>
<td></td>
<td>Parallel, continuous, reflective</td>
<td>Glacio-lacustrine to lacustrine sedimentation</td>
</tr>
<tr>
<td>U11b</td>
<td></td>
<td>Semis-transparent</td>
<td>Proglacial debris-flows from the Nyon readvance</td>
</tr>
<tr>
<td>U10</td>
<td></td>
<td>Parallel, continuous, reflective</td>
<td>Glacio-lacustrine, proglacial sedimentation</td>
</tr>
<tr>
<td>U9b</td>
<td></td>
<td>Semi-transparent</td>
<td>Proglacial debris-flows from the Coppet readvance</td>
</tr>
<tr>
<td>U8</td>
<td></td>
<td>Parallel, continuous, reflective</td>
<td>Glacio-lacustrine, proglacial sedimentation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Chaotic with weak oblique stratifications dipping downflow</td>
<td>Foresets deposited at the front of the receding glacier</td>
</tr>
<tr>
<td>U6 (facies a and b)</td>
<td></td>
<td>U6a: reflective, continuous, subparallel, dome-shaped</td>
<td>U6a: esker</td>
</tr>
<tr>
<td></td>
<td></td>
<td>U6b: semi-transparent to chaotic</td>
<td>U6b: lateral evolution of eskers into fine grained deposits</td>
</tr>
<tr>
<td>U4-U5</td>
<td></td>
<td>Chaotic to semi-transparent with some isolated, strong, continuous reflections</td>
<td>Subglacial melt-out till deposition. Strong reflections indicate readvances of the glacier associated to sediment compaction and shearing</td>
</tr>
<tr>
<td>U3</td>
<td></td>
<td>Semi-transparent with weak, low-frequency, parallel reflections</td>
<td>Sediments deposited in a subglacial lake at the onset</td>
</tr>
<tr>
<td>U2 (facies a and b)</td>
<td></td>
<td>U2a: reflective, subcontinuous, subparallel, dome-shaped in places</td>
<td>U2a: sands and gravels deposited at the bottom of the glacial valley at the onset of the deglaciation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>U2b: semi-transparent to chaotic</td>
<td>U2b: finer fluvio-glacial sediments or till</td>
</tr>
<tr>
<td>U1</td>
<td></td>
<td>Semi-transparent</td>
<td>Glacial sediments older than the würmian glacial maximum</td>
</tr>
<tr>
<td>U0</td>
<td></td>
<td>Sub-horizontal, very reflective, continuous, with sharp lateral truncation</td>
<td>Molasse bedrock</td>
</tr>
</tbody>
</table>

Table 11-1 Seismic stratigraphy of Western Lake Geneva, with illustration and description of the seismic facies.

Inferred. Given the high resolution of the seismic data and the resulting thinness of the reflections, the profiles are always presented with a certain vertical exaggeration. On longitudinal profiles (parallel to the lake axis), glacier flow is always from right to left. The seismic stratigraphy is summarized in Figure 11-2, which represents a synthetic longitudinal section passing through the glacial valley talweg. Within a given unit, distinct seismic facies were named with a letter suffix (e.g. 2a, 2b) and stratigraphical subdivisions with a numerical suffix (e.g., 8.1, 8.2).
11.2 Molasse bedrock (U0)

U0 is the lowermost unit, recognizable over the entire area and consists of continuous sub-horizontal reflections of high amplitude and low frequency. The upper boundary is marked by a sharp truncation of the internal reflections, particularly well visible on seismic lines transverse to the lake axis, where the truncated reflections form large steps. Unit 0 is deeply eroded by a large valley following the lake axis.

These characteristics are typical of the eroded Molasse basement, formed by marls and sandstones. The isochron map of this unit upper surface (Fig. E-2) reveals that the valley is shallowing and narrowing towards the south. It forms part of an anastomosed network of subglacial channels mapped in figure 11-3.

The Western Lake Geneva axis follows roughly a broad Molasse syncline axis, evidenced by the Molasse bedrock layers dipping towards the lake centre on both the western and eastern shores (Fig. D-3). As the data are not depth-converted, the increasing thickness of low-velocity layers (water and Quaternary deposits) towards the lake centre may lead to a small exaggeration of the Molasse dip. As already noted by Girardclos (2001), the normal faults indicated by Vernet and Horn (1971) do not exist. At the latitude of Céligny, a tectonic accident oriented E-W (100°) crosses several seismic lines (Fig. D-6). It is interpreted as the prolongation of the La Faucille strike-slip fault (Meurisse et al., 1971), one of the numerous 100°-oriented, dextral strike-slip faults affecting the Jura limestones (Fig. 11-4). The dip of the layers around the fault indicates a vertical component, with relative elevation of the southern block (Fig. D-6).

The transition from the Grand Lac to the Petit Lac is marked by a rapid narrowing and shallowing of the glacial valley, passing from more than 12 km off Rolle to ca. 4 km off Nernier. In the Petit Lac, the bedrock valley shallowing initiated in the Grand Lac continues, passing from 360 ms off Yvoire to 120 ms near Geneva (324 to 108 m asl for a velocity of 1800 m/s). At the deepest points of the glacial valley, the seismic signal strength is highly attenuated by the overlying deposits, making the bedrock depth uncertain and potentially more important.

In the northeastern part of the Petit Lac, the valley presents an asymmetric V-shaped cross-section with a steeper eastern flank. This asymmetry, also present in the Grand Lac (Dupuy, 2005), is enigmatic. It does not result from the Coriolis effect, which predicts an ice/melwater flow deviation towards the right for the northern hemisphere and a higher erosion of the western flank. It is neither explained by a structural control, as the broad anticline centred on the Petit Lac is symmetrical.

The cross-section remains stable until Coppet, where the main valley is deported towards the east while a narrow, 4.5 km long, sub-parallel secondary valley appears on the western flank of the main valley.
This secondary valley, which narrows rapidly and ends abruptly after 3 km, may have two origins. First, it may have been carved by water and ice during a glacial cycle older than the Würmian glaciation. A similar secondary valley is visible on Lake Neuchâtel seismic sections and cores have shown that it was created and filled prior to the Würm glacial cycle (Clerc, 2006; A. Pugin, pers. comm.). Alternatively, it could be part of the Rhône glacier tunnel channel network and full of water during the last glaciation. Its abrupt termination is typical of tunnel channels (e.g., Huuse & Lykke-Andersen, 2000). In any case, stratigraphical relationship and velocity analysis (section 11.6) indicate that the secondary valley contains sediments older than the main valley infill; it is therefore named “Coppet paleo-valley” (Fig. E-2).

Off Anières, the main valley turns west and a shallower branch continues towards the present-day Corsier bay, east of the Hauts Monts bedrock knob (Girarde los, 2001). In this same area, the valley eastern flank presents truncation of older Quaternary sediments revealing the presence of an “Anières paleo-valley” (Fig. D-7). Towards the south, this valley could be connected to a valley to the southeast of the Aire plaine, recognized on cores and seismic (A. Pugin, pers. comm.).

The relation with the Grand Lac morphology (Dupuy, 2005) is visible on the general bedrock map of Western Switzerland (Fig. 4-1). Between Cully (x = 546.0) and Buchillon (x = 522.0), the Grand Lac also presents a V-shaped and asymmetrical cross-section with a steeper southern flank. Between Buchillon and the Petit Lac, the bedrock gets shallower and loses its V-shaped cross-section. “Channels” are shallow, poorly marked and difficult to follow between seismic sections.

On the Grand Lac map of “glacial channels” (Fig. 68 in Dupuy, 2005), the talweg of the main glacial valley is called “channel south”. As the talweg is by definition a line without lateral extent or limits, the term “channel” is not appropriate and the width indicated on Dupuy’s map is subjective. Off Cully, between x coordinates 541.0 and 546.0, the valley northern flank is cut by a secondary channel (“channel North”), connected at its two extremities to the main talweg. Contrary to Dupuy, our examination of his seismic lines and of our own lines do not indicate a northern channel between Lausanne and Buchillon. This channel does exist however west of Buchillon and constitutes the continuation of the Petit Lac.

Compared to the previous bedrock map of Vernet and Horn (1971), our denser coverage and deeper penetration reveal a much sharper relief and a deeper erosion in the centre of the valley (Fig. E-2). It completes the work of (Dupuy, 2005) for the Grand Lac. The difference between the bedrock and lake bottom arrival times results in an isopach map of the Quaternary sediments (Fig. E-3).
11.3 Quaternary seismic units (U1-U14)

11.3.1 General presentation

The Quaternary infill of the Petit-Lac consists mostly of glacial sediments, topped with a relatively small thickness of glacio-lacustrine and lacustrine sediments. The subdivision of the glacial sedimentary record into seismic units (U1-U7, U9, U11) is based on strong and continuous reflections interpreted as major erosion surfaces. This subdivision is particularly evident in the northern part of the Petit-Lac, where units are laterally continuous and the bedrock valley is straight and wide (Fig. D-3). In the central part of the Petit-Lac (between Coppet and Versoix), the path of the bedrock valley is more curved and the geometrical relationship between units is more complex, with rapid lateral variations of facies and truncation of older units by younger erosion surfaces. The glacial record south of Versoix is more homogenous: seismic units are less numerous because of the increasing truncation of seismic units towards the south, and the seismic interpretation is more subjective, due to the presence of gas.

11.3.2 U1 - ancient glacial sequence

U1 consists of all the sediments deposited between the bedrock (U0) and U2 and presents a variable facies ranging from semi-transparent to chaotic. Its main accumulation is the complete infill of the Coppet paleo-valley (Fig. D-4). It is also present in the Anières paleo-valley, where it is eroded laterally by the flank of the last glacial valley, and in the southern part of the lake, at the bottom of the main valley (Fig. D-5).

Several elements suggest that U1 is a sedimentary sequence older than the last glacial cycle: its presence in the Coppet secondary valley, its truncation by the main valley along the eastern shore near Anières and its higher acoustic velocity (see section 11.6). In Switzerland, such deposits are usually attributed to the Rissian (e.g., Amberger, 1978) or early-Würm (e.g., van der Meer, 1982) glacial cycle.

11.3.3 U2 - subglacial gravels

U2 lies directly over the Molasse bedrock or over U1, in the deepest northern part of the Petit-Lac valley (Fig. E-4). Two seismic facies are present within this unit (Table 11-1): reflective with sub-parallel/cross-stratified internal reflections (facies 2a) or semi-transparent (facies 2b). U2 is limited upward by a very strong reflection which truncates its internal stratifications. On transverse sections, this upper surface is undulating and may present domes, which contain backsets visible on longitudinal sections.

Based on the reflectivity of facies 2a and on the dome structures, this unit is interpreted as gravels deposited under the glacier by meltwater. The domes have a relatively small longitudinal extension and could correspond to short eskers. In Lake Zürich, gravels were also reported at the bottom of the sedimentary sequence (Lister, 1984b). The semi-transparent facies 2b could correspond either to massive gravel/sand deposits also deposited by meltwater, or to intercalations of till deposited by ice. This unit is the first deposit of the last glacial cycle and the following units continue to fill progressively the glacial valley.

11.3.4 U3 - subglacial lake sedimentation

The base of U3 is bounded by a strong reflection, indicating a marked change in acoustic properties from U2 to U3 related to a density decrease. U3 presents a typical stratified facies, with reflections of medium energy and low frequency. The internal sub-parallel reflections are generally conformable with the underlying deposits, but onlap on the glacial valley borders and on the irregular bumps of U2 (Fig. D-3). Thanks to its typical stratified facies and its wide extension (Fig. E-5), this unit constitutes a “marker” level allowing a reliable correlation of the northern and southern part of the Petit-Lac.

U3 is interpreted as subglacial lacustrine deposits. The exact sedimentary process is unknown but the stratified nature of the sediments might suggest decantation and/or turbidites. The establishment of a water body between the glacier and its bed marks the onset of ice thinning during the deglaciation.


11.3.5 U4 and U5 - till

U4 and U5 are acoustically semi-transparent to chaotic. Towards the south, the facies of U4 passes from semi-transparent to reflective, with an increase in the density of diffractions and with short reflections. These units are bounded by continuous reflections with a high reflectivity on the valley borders, which decreases towards the valley center. The top of these units is deeper in the centre of the lake (Fig. E-6).

U4 and U5 are interpreted as tills deposited under a floating glacier. Their lithology is probably similar to the thick clayey till units cored in Lake Zürich (Lister, 1984b). The strong reflections separating them are interpreted as compaction levels formed by the re-coupling of the glacier to its bed during readvances punctuating the general retreat.

11.3.6 U6 - eskers

North of Hermance, U6 consists mainly in a thin reflective level associated to numerous diffractions immediately below (facies 6a, Fig. D-3). This facies is well developed near the lake borders, and the reflectivity decreases towards the lake center, where only a few diffractions are present. The top of this unit is deeper at the centre of the valley (Fig. E-7). Along the lake borders, short, low-relief, reflective ridges are present (Fig. E-8).

Towards the south, the ridge situated along the western shore becomes progressively more prominent. Between Hermance and Anières, this western ridge becomes very developed and a second ridge appears to the east, at the centre of the valley (Figs. D-4 and E-8). Both ridges are parallel, rectilinear and present a SW-NE elongation over ~2.5 km, well visible on time-slices (section 11.5). These two ridges are still visible today in the lake bottom morphology (Sastre, 2004). In cross-section, each ridge contains strong reflections forming a concentric antiform structure (Fig. D-8). This internal structure is disrupted by unconformities. Downflow, between Anières and Versoix, the ridges progressively widen, lose their reflectivity, and only a few undulating reflections subsist, corresponding to the internal unconformities seen upflow. South of Versoix, reflections disappear completely and the facies is semi-transparent (facies 6b).

Between the two ridges, transverse seismic sections show a 150-200 m wide trough (Fig. D-8), with a height and length corresponding to the rectilinear segments of the eskers. This through has steep flanks, a flat bottom and a slightly convex-upward long-profile. The trough flanks erode laterally the esker sediments, particularly along the eastern esker. Its infill presents a faintly stratified to semi-transparent seismic facies.

The ridge morphology and the stratified reflective facies clearly point to eskers made of gravels and sands deposited by subglacial meltwater circulation, like previously described in former seismic-stratigraphic studies (e.g., Mullins et al., 1996; Pugin et al., 1999; Beck et al., 2001; Noormets & Floden, 2002). An englacial or supraglacial origin for the esker is unlikely, given the absence of deformation in the deposits. Unconformities within the esker must correspond to erosive events due to pulses of meltwater. North of Hermance, the predominance of flat diffraction levels and the poor development of eskers suggests that meltwater flowed in very low, lens-shaped conduits. This variation in the conduit shape towards the ice front suggests that eskers can only form under a relatively low ice pressure, near the ice front.

The western esker presents downflow variations in size, shape and facies (Fig. D-8). The sedimentary sequence becomes broader and higher, the cross-section evolves from a single crest to a multi-crested esker and the acoustic facies evolves from reflective to semi-transparent. All these tendencies indicate an increase in the conduits size towards the ice margin and the evolution of the esker into an outwash fan with deposition of finer sediments due to flow expansion and deceleration when the meltwater enters the proglacial water body (Fig. 11-5). Such sedimentary bodies are known as “esker deltas” (Banerjee & McDonald, 1975; Rust & Romanelli, 1975; Sharpe, 1988; Warren & Ashley, 1994), and were previously imaged by seismic reflection in Lake Annecy (Van Rensbergen, 1996). Because of the meltwater high sediment load, the incoming flow is denser than the proglacial lake water, leading to underflows (Brennand, 2000).

Three sedimentary sequences constitute the western esker, well visible on the profile shot longitudinally above
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the esker axis (Fig. D-9). This profile is not located exactly on the esker crest, but cuts it slightly obliquely in the northern part (Fig. 11-6). Sequences thicken downflow and are separated by strong continuous reflections. Their base evolves downflow from slightly erosive to conformable. On transverse sections, the recognition of these sequences is not straightforward because of their very small lateral extent. In the northeastern part of the western esker, the sequence limits correspond to angular unconformities within the gravel body (e.g. line 121 on Fig. D-8). Downflow, these unconformities evolve into well-defined, conformable, undulating, multi-crested reflections separating semi-transparent sequences (see southernmost sections on figure D-8).

The eastern esker is single-crested, narrower and slightly shorter than the western esker. Its cross-section is ~200 m wide and presents a ~35 m high single dome with a steep northwestern flank and a gentler southeastern flank. From the northeast, the esker longitudinal section (Fig. D-10) grows very rapidly and presents a steep upflow extremity. The reflective stratified facies and the sharp dome shape remain stable over the first 2 km, then the reflectivity decreases strongly and the typical dome shape disappears. The longitudinal section (Fig. D-10) reveals that this rapid transition results from an abrupt increase in the conduit height, due to a downward step at the base of the conduit.

Both eskers present generally stratified sediments but acoustically chaotic or semi-transparent ones are also present. The latter may correspond to unstratified or poorly stratified matrix-supported gravels (e.g., Syverson et al., 1994; Fig. 11.32 in Benn & Evans, 1998) or to diapiric intrusion of underlying fine sediments due to the differential ice pressure between the esker sides and the esker centre (Banerjee & McDonald, 1975). Both stratified and unstratified esker infills were reported from esker outcrops (Fig. 11-7).

The longitudinal trough between the western and eastern eskers (Fig. D-8) clearly involves erosion after the esker formation, as it truncates the esker layers. Its

Fig. 11-6 Schematic drawing presenting the spatial relationship between the western esker and the longitudinal seismic section AG1_JF_03.c2a. The esker is composed of two crests indicating a migration of the system towards the west.

Fig. 11-7 Two examples of contemporaneous esker infills in Alaska (from Syverson et al., 1994): a) esker “Anticlinal” bedding of sands and gravels; b) Massive subglacial infill of a 3 m wide subglacial tunnel.
flat bottom and its steep walls present similarity with the tunnel channels formed under the Laurentide ice sheet. However, the trough cannot be followed outside of the area containing the two eskers. Besides, meltwater channel formation after the eskers is unlikely, as many studies show the inverse sequence (Pugin et al., 1999; Fisher et al., 2002; Sjogren et al., 2002; Fisher et al., 2005). Given the rectilinear path of this trough and its geometrical association with eskers, erosion by ice is more likely. During the formation of the eskers, the elongated area between them was filled with ice from the overlying glacier. Following the esker infill with sands and gravels, a small glacial readvance has eroded the esker borders and probably their upflow extremity. Such reworking has often been described from outcrops with deformation ranging from superficial shear planes (e.g. Gorrell & Shaw, 1991) to strong fracturation, deformation and displacement (e.g. Knudsen, 1995). Finally, the ice thinned and became uncoupled from the bed, leading to the infill of the trough with post-esker deposits, probably consisting of fine-grained fan sediments (Sharpe, 1988).

11.3.7 U7 - ice retreat deposits

U7 lies on top of the esker-fan system and its thickness decreases towards the south (Fig. 11-2). North of km 10 on figure 11-2, its facies is semi-transparent, similar to that of U4 and U5. South of km 10, its facies becomes chaotic, with discontinuous reflections. Weak reflections dipping towards the south are present between km 12 and 15 (Figs. 11-2 and D-11).

The high stratigraphic position of U7, immediately below the stratified lacustrine sediments of U8, suggests a deposition near the ice front during ice retreat. This hypothesis is supported by the internal south-dipping reflections, which correspond probably to foresets deposited at the ice margin by meltwater and/or avalanching of proglacial sediments bulldozed by the oscillating ice front (Fig. D-11), and by the presence of faint stratifications at the top of U7 in the central part of the Petit-Lac (Fig. D-12).

11.3.8 U8 - glacio-lacustrine

U8 facies contrasts clearly with U7 and presents conformable, very continuous, high-frequency, medium- to high-amplitude reflections.

This facies is typical of stratified sediments deposited in a low-energy water body and U8 is interpreted as the first glacio-lacustrine unit, constituted of small debris-flows, turbidites and underflow sediments coming from the base of the active glacier front, grading progressively to lacustrine deposits as the glacier receded (Girardclos, 2001; Baster et al., 2003; Girardclos et al., 2005). If we consider individual reflections as isochrones, the transition from U7 to U8 is slightly time-transgressive because the lowermost reflections of U8 are present only in the southern part of the lake. This time-transgressive facies change marks the retreat of the ice front in a proglacial lake.

Along the eastern shore, between Hermance and Tougues, the glacio-lacustrine sedimentation started earlier, indicating either the existence of a subglacial cavity or the earlier retreat of the ice in this zone due to the meltwater input from a paleo Hermance river (Figs. D-13 and E-9).

The difference between the base of the glacio-lacustrine sediments and the Molasse bedrock provides an isopach map of the glacial sediments, i.e., of units U1-U7 (Fig. E-10).

11.3.9 U9 - Coppet readvance

U9 can be separated in three sub-units: 9a, 9b and 9c, well visible on the profile presented in figure D-14.

U9a

U9a is a glacio-tectonic unit characterized by thrust folds affecting the sediments of U9b, U8 and the upper part of U7 (Figs. D-14 and D-15). The thrust faults dip towards the NE with an apparent dip of ~8°, and join downwards a décollement plane slightly dipping towards the NE. The horizontal wavelength of the disrupted folds is ~55 m and the amplitude is ~3.8 m (4.8 ms twt). Deformation increases progressively towards the NE because the glacio-lacustrine stratifications in the thrust slabs get more discontinuous and perturbed until they finally lose their stratified facies some 800 m away from the deformation front. However, the contact between the deformed lacustrine sediments and the overlying semi-transparent sediments remains visible. Deformation of proglacial sediments is also visible on transverse profiles (Figs. D-3 and D-16), where U8 glacio-lacustrine sediments are folded and faulted at the centre of the valley but are not or little perturbed towards the valley borders. The deformation is thus stronger at the centre of the section and indicates a convex deformation front in plan view.

These observations indicate a readvance of the Rhone glacier. During the latter, previously deposited sediments (mainly glacio-lacustrine) were bulldozed by the ice front, leading to the development of a glacio-tectonic thrust system associated to a basal décollement plane. These characteristics are typical of a push moraine, i.e., a glaciotectonic ice-marginal moraine (van der Wateren, 1994, 1995; Bennett, 2001). The compression of the proximal zone led to the progressive extension of the basal décollement (Boulton et al., 1999).

The basal reflection of the décollement plane can be followed towards the north, where it is interpreted as a compaction level corresponding to the glacier base. The map of this surface (Fig. E-11) reveals a “trowel” shape: deeper at the lake centre and shallower towards
the south. The isopach map of the Coppet readvance (Fig. E-12) shows the large volume of sediments affected.

The northward facies transition from stratified to discontinuous and chaotic reflects an increasing deformation. Glacio-lacustrine sediments are transformed into a glaciotectonite evolving later to a deformation till, the limit between these two terms corresponding to the loss of structural characteristics of the parent material (Benn & Evans, 1998). Given the limited horizontal resolution of the airgun seismic data for the Coppet thrust moraine (~12 m for a depth of 40 m) and the trace spacing of 3.5 m, a block of several meters can remain undetected. Therefore, a chaotic to transparent seismic facies does not necessarily indicate completely homogenized sediments, but might also contain relatively large blocks of undisrupted sediments (Schnellmann et al., 2005).

**U9b**

U9b is a semi-transparent, unstratified wedge intercalated between the stratified units U8 and U10 without perturbation of the underlying sediments (Figs. D-14 and D-18). Towards the SW, the wedge pinches out; towards the NE, it is separated from U9a by a thrust fault. Its length is about 1800 m and its thickness reaches 10 ms (= 8 m for a velocity of 1600 m/s) in the up-glacier part. The lower contact is sharp and conformable; the glacio-lacustrine layers show no erosion. They are very slightly and irregularly bended, due to the wedge sediment loading. The upper contact is also conformable. On transverse profiles beyond the deformation front (Fig. D-17), U9b pinches out towards the valley borders. On transverse profiles crossing the curved deformation front (Fig. D-16), U9b is restricted to the sides of the valley, as the central glacio-tectonized part is up-lifted.

The semi-transparent, unstratified facies of U9b is interrupted in places by a coarsely stratified and highly reflective facies, well visible at the distal extremity of the wedge in figure D-18. Reflections are sub-horizontal or arranged in foresets dipping apparently up-glacier. This reflective facies is truncated by the transparent facies, and must therefore be older. This stratified facies is also present on the transverse profiles (Fig. D-17), in lateral contact with wide channels filled with the transparent facies, indicating an erosion by the channels and a subsequent infill by acoustically transparent material.

U9b is interpreted as a wedge of glaciogenic debris-flows originated from the upper surface of the bulldozed sediments forming lenses of unstratified sediments beyond the deformation front. The large longitudinal extension of this unit results from the ability of debris-flows to travel for long distances on very gentle slopes as low as 0.1° (Mahgoub, 1998). This sub-unit corresponds to the “Coppet” till tongue of Moscariello et al. (1998). Near the glacio-tectonic deformation front, debris-flows filled the lateral depressions on both sides of the material uplifted by the glacier bulldozing. As the glacier advanced, the proximal part of U9b was progressively incorporated into the push-moraine (upper part of U9a).

The parts of U9b presenting a reflective facies are interpreted as remnants of material deposited as an ice-contact subaqueous fan by underflow meltwater streams coming out from subglacial or englacial meltwater conduits at the glacier front (Eyles et al., 1985). Then, the sedimentation switched to more erosive channelized gravity-flows (e.g., Cai et al., 1997; Lonne & Syvitski, 1997) leading to unsorted and unstratified deposits. This sedimentary process is supported by the faint convolved reflections seen on the longitudinal sparker profile presented in Fig. D-18, interpreted as gravitational flow folds (Hart & Roberts, 1994). These debris-flows eroded the subaqueous fan, leaving only small stratified inter-channels. The transition from stratified deposits to massive deposits is probably the result of sediment instabilities on the flanks of the pushed sediments during the readvance.

A similar scenario has been proposed by Bennett et al. (2002) for a recent glacio-lacustrine sediment accumulation in the Copper River Basin (Alaska). At first, a proglacial fan builds up through the accumulation of gravels and sands deposited directly from meltwater and through sediment-density currents. Subsequently, mud-rich sediments are deposited through channelized bottom currents and sediment-density flows. The mud-rich sediments erode the previously deposited gravels and sands. Cai et al. (1997) also reported the formation of channelized gravity flow deposits (probably high-density turbidites or debris flow deposits), originating from morainal banks in a fjord.

**U9c**

Unit 9c is located to the north of unit 9b, behind the thrusted material. Its facies is chaotic and difficult to differentiate from unit 9b, but the disappearance in U9c of the U7/U8 boundary reflection constitutes a good criterion. On lateral longitudinal lines, the limit between 9b and 9c is also marked by a strong oblique reflection dipping up-ice with an angle of 3 to 5°.

Stravers and Syvitski (1991) reported a similar angle (~5°) for the ice-sediment contact in the Cambridge Fjord, Baffin Island. Therefore, these oblique reflections are likely to represent the Coppet readvance ice front. U9c is interpreted as ice-marginal sedimentation infilling the space left between U9b and the ice front when the glacier receded.

The maximum position reached by the glacier during the Coppet readvance, indicated on the longitudinal profiles, is also inferred from the proximal flank of the smooth and low transversal bank visible in the lake bottom morphology (Fig. D-14), and from the recognition on the DEM of onshore fronto-lateral moraines (Fig. 7-18). The
(smaller and lower lake bottom moraine compared to the terrestrial moraines is the consequence of the ductility of the water-saturated sediments during its formation, and of the thick (~20 m) cover of younger glacio-lacustrine and lacustrine sediments.

U9c also extends over the push-moraine as a thin transparent layer over the external piggy-back structures (Fig. D-18). No glacio-tectonism is perceptible on the seismic data, indicating that this tiny part of the deposition occurred at the maximum of the readvance or after the readvance.

11.3.10 U10 - glacio-lacustrine

U10 presents a stratified, continuous, conformable facies and is interpreted as a second glacio-lacustrine unit deposited between two glacial readvances.

11.3.11 U11 - Nyon readvance

U11 presents the same characteristics as U9 and is therefore interpreted as a second positive oscillation of the receding ice front, corresponding to the “Nyon” readvance previously reported by Moscariello et al. (1998). Like the Coppet readvance, it is subdivided into three units: U11a (push-moraine), U11b (proglacial debris-flows) and U11c (ice-marginal sediments deposited during ice retreat). This readvance is well illustrated by the airgun 15-inch1 longitudinal profile ia4 (Fig. D-19). From south to north, the strong reflection at the base of U10 gets increasingly thrusted over a length of 850 m. The thrust faults have an apparent dip of ca. 8° and affect both U9c and U10. Behind the thrust zone, a 250 m wide basin is marked by another strong reflection, which becomes suddenly horizontal at the basin northern margin.

Cross-sections (Fig. D-20a) and the isochron map (Fig. E-13) show that the glacio-tectonized unit is thicker at the centre and its base is V-shaped. A more distal cross-section (Fig. D-20b) shows that thrust faults dip towards the lake centre, indicating a convex deformation front in plan view and a compression not only in the lake axis direction but also towards the lake shores.

In the valley centre, the pushed sediments form a positive relief. During its formation, slope instabilities in unconsolidated material led to the triggering of debris-flows (11b) which accumulated in the lateral depressions, mainly on the western side (Figs. D-20 and E-14). At the centre of the lake, U11b is very thin (Fig. D-19). On the valley sides, the lacustrine sediments of U10 are buried by the debris-flows prior to deformation, whereas they are directly folded and thrusted at the valley centre.

The Nyon push-moraine was previously interpreted as a slump (slump_F) by Baster (2002). This misinterpretation was the result of the echosounder seismic source low penetration (25 ms) and of the similar seismic facies of slumps and glacio-tectonized sediments (Syvitski, 1991).

The stratigraphical relationship between the Coppet and Nyon readvances is illustrated in figure D-21 and the morpho-tectonic elements are mapped in figure E-15.

11.3.12 U12-U14 - glacio-lacustrine to lacustrine

Units 12 to 14 present a stratified, continuous, conformable facies and are interpreted as glacio-lacustrine to lacustrine sediments deposited after the Nyon readvance. This subdivision in three units results from the very-high resolution seismic stratigraphy of these upper deposits (Girardclos et al., 2003; Girardclos et al., 2005), summarized in figure 11-8.

On airgun data, the most remarkable reflection in these three units is the seismic reflection 14 of Girardclos et al. (2005), located in U13 and assigned to the Younger Dryas / Preboreal transition (Fig. 11-8). This level is very reflective and continuous and is associated with three
Fig. 11-9 Time-slices through the Petit Lac pseudo-3D survey, envelope attribute. a) Inline 120 envelope attribute showing the depth of the time slices. For inline location, see the 40 ms time slice. b) Time slices ranging from 40 to 260 ms, with an increment of 20 ms. Annotations: lb = lake bottom, l = lacustrine, gl = glacio-lacustrine sediments, vb = glacial valley border, m = molasse, e = esker. See text for explanations.
slumps (Fig. E-16): two large ones north of Versoix and off Tougues and a smaller one off Hermance. The first two have already been recognized by Girardclos et al. (2005) but the Hermance slump has not yet been reported. The occurrence of three slumps at the same stratigraphic level probably indicates that these deposits were triggered by an external mechanism, probably an earthquake. Above this reflective level, sections transverse to the lake show a transparent unit filling in onlap the centre of the lake (Fig. D-12). This corresponds probably to large underflows (turbidites) triggered either by the slumps or by the hypothetical earthquake.

Finally, the present day lake bottom (Fig. E-17) consists in a strong positive reflection. One to three low-frequency, very-low-amplitude reflections are generally present just above the lake bottom. They correspond to artifacts produced by low-pass filtering during the processing.

11.4 Isopach map of the stratified sediments

The difference between the base of the glacio-lacustrine sediments and the lake bottom provides an isopach map of the undisturbed stratified deposits (glacio-lacustrine, lacustrine and till-tongues), represented in figure E-I8. It is highly influenced by the pre-existing relief, with higher sediment accumulation in the deeper zones and less accumulation on the margins and the Hauts-Monts area. The thickest sediments are located in the vicinity of Hermance with a thickness of 60 ms. This results from the deposition of older stratified sediments in this zone (Fig. D-13).

The general trend is a thinning of these deposits towards the North. This tendency results from the progressive retreat of the ice from the basin, leading to the earlier apparition of the glacio-lacustrine sedimentation in the south. Two important steps corresponding to the glacio-tectonic fronts of the Coppet and Nyon readvance affect this thinning. North of each tectonic front, the glacio-lacustrine sequence deposited before the readvances has been tectonized so as to loose any stratification. Therefore, only the sediments deposited after the readvances remain stratified.

The isopach map also shows a thinning of the sediments away from each glacio-tectonic front, indicating that the main sediment input came from the glacier front and the associated glacio-teconic front. The wedge shape of the debris-flows (or till-tongues) triggered from the glacio-tectonic front are the primary reason of the whole glacio-tectonic sequence thinning.

West of the Hauts Monts area, the amount of stratified sediments is reduced, due to the Holocene erosion by lacustrine currents (Girardclos, 2001).

11.5 Envelope time-slices

Time-slices are horizontal cuts through a 3D seismic cube, which offer an alternative view of the seismic data, highlighting the lateral continuity of seismic units. In true 3D cubes, as those used in the oil industry or in the resent work of Dupuy (2005) on Eastern Lake Geneva, time-slices are usually cut through the original seismic data. With pseudo-3D cube used in the present study, time slices through the original data result in meaningless images because the inlines are too widely spaced to follow the phase between them. Besides, the short wavelength of the data makes it very sensible to the small time shifts between inlines, resulting from small differences in the airgun and streamer depths.

To get around this difficulty with the original seismic data, we generated the envelope attribute to obtain meaningful time-slices. The envelope attribute represents the total instantaneous energy of the seismic data and its magnitude is of the same order as that of the input traces. Its value is proportional to the acoustic impedance contrast, hence the reflectivity, and varies approximately between zero and the maximum amplitude of the trace. It is independent of the phase and represents the combined response of several interfaces, making it useful to characterize and differentiate the main seismic units (Taner, 2001).
Time-slices through the envelope attribute provide therefore images where the reflective units (bedrock, gravels) contrast clearly with the less reflective ones (till, sands, silts). Figure 11-9 shows a sequence of time-slices from 40 to 260 ms with a 20 ms increment, and permits the following observations:

20-80 ms: These three first slices pass progressively from the transparent water (at the center) to the reflective lacustrine sediments (on the borders). The southern part of the survey is more noisy (yellowish) because the original 2D surveys were acquired in wavy conditions.

100 ms: The slice cuts through the highly-reflective lacustrine sediments.

120 ms: The slice cuts through the poorly reflective glacio-lacustrine sediments, which contrast strongly with the glacial valley borders, highlighting its “S” shape. The eastern valley border corresponds to the Molasse erosion, while the western border corresponds to the old glacial deposits of U1, eroded and compacted during the last glacial cycle.

140-180 ms: These slices are located in the esker zone, which gets progressively more reflective. Eskers coarse deposits appear as SW-NE oriented reflective stripes corresponding to the centre of the meltwater tunnels.

200-260 ms: At this depth, the acoustic signal is too attenuated to interpret the time-slices correctly. The northeastern part of the cube, corresponding to survey AG1_AP_00 is less noisy that the southwestern part, corresponding to survey AG1_JF_02 (Table 10-1).

## 11.6 Interval velocities

In addition to the stacking velocity analyses included in the processing flow, three seismic sections were analysed more precisely to obtain interval velocities (i.e., the sediments true velocity) and to interpolate velocity sections (see the script velmod.sh in Appendix F). Ideal velocity estimates should be done on horizontal and continuous reflections, as dip affects the stacking velocity. The southern end of longitudinal line AG1_JF_05.11 (Fig. 11-10) satisfies these conditions. Based on a selection of 23 CDPs, we obtained the interval velocities for the main units (Table 11-2).

The water velocity of 1430 m/s corresponds to the theoretical freshwater velocity for a temperature of 6°C and a depth of 50 m (Coppens, 1981). For the glacio-lacustrine sediments, the velocity of 1554 m/s is rather low and indicates a high water content. It corresponds to the values reported previously in Lake Geneva – 1450-1650 m/s (Scheidhauer, 2003), 1490-1700 m/s (Girardclos, 2001) – and in other perialpine lakes – 1550-1650 m/s (Finchek et al., 1984). Deeper, the velocities increase progressively down to unit U1 which presents a high velocity of 2625 m/s, indicative of more compact sediments.

Because of the streamer small length with respect to the depth investigated, the velocity values present a certain variability, indicated in Table 11-2 by the standard deviation. This variability increases with depth, because of the difficulty to pick the correct velocities on the semblance panels, and is accentuated by the small thickness of the deeper layers (U2 and U3).

The esker sediment velocity were analyzed on the line AG1_JF_03.c2a central part (CDP 416 to 829), located right over the main esker body and presenting sub-horizontal reflections. A selection of 11 CDPs indicates a mean velocity of 1789 m/s.

We finally analysed the velocities of the 5-inch transversal profile AG5_AP_96.ag08 to see whether the Coppet secondary valley infill presented a higher velocity than the main valley infill (Fig. 11-11). If so, it could indicate overcompacted sediments older than the main valley sediments, maybe pertaining to a previous glacial cycle. Given the changing dip along the section, the standard deviation is high and velocities are not accurate. However, the secondary valley has a slightly higher velocity, suggesting a higher compaction and therefore a probable deposition during a previous glacial cycle. For the Molasse bedrock, we obtain a velocity of ca. 3000 m/s, probably underestimated because of the possible confusion with the lower-velocity U1 and U2 units.

<table>
<thead>
<tr>
<th>unit</th>
<th>mean interval velocity (m/s)</th>
<th>standard deviation (m/s)</th>
<th>average thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>water</td>
<td>1430</td>
<td>4</td>
<td>69.7</td>
</tr>
<tr>
<td>(glacio-)lacustrine</td>
<td>1554</td>
<td>22</td>
<td>35.7</td>
</tr>
<tr>
<td>U4-U7</td>
<td>1817</td>
<td>77</td>
<td>112.4</td>
</tr>
<tr>
<td>U3</td>
<td>1995</td>
<td>224</td>
<td>31.4</td>
</tr>
<tr>
<td>U2</td>
<td>2625</td>
<td>363</td>
<td>29.4</td>
</tr>
</tbody>
</table>

Table 11-2 Interval velocities for the southern end of line AG1_JF_05.11. Note the progressive velocity increase with depth and the high-velocity of unit U1.

<table>
<thead>
<tr>
<th>unit</th>
<th>mean interval velocity (m/s)</th>
<th>standard deviation (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>water</td>
<td>1427</td>
<td>7</td>
</tr>
<tr>
<td>(glacio-)lacustrine</td>
<td>1512</td>
<td>23</td>
</tr>
<tr>
<td>esker</td>
<td>1789</td>
<td>120</td>
</tr>
</tbody>
</table>

Table 11-3 Interval velocities for the central part of line AG1_JF_03.06, located over the esker deposits.

<table>
<thead>
<tr>
<th>unit</th>
<th>mean interval velocity (m/s)</th>
<th>standard deviation (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>main valley infill (below eskers)</td>
<td>2202</td>
<td>227</td>
</tr>
<tr>
<td>Coppet secondary valley infill (below eskers)</td>
<td>2501</td>
<td>532</td>
</tr>
<tr>
<td>Molasse bedrock</td>
<td>3006</td>
<td>482</td>
</tr>
</tbody>
</table>

Table 11-4 Interval velocities for line AG5_AP_96.ag08, comparing the velocities of the main and secondary valley infills.
The sediment velocities presented here are in agreement with those reported in previous studies. As far as the Molasse bedrock is concerned, sonic velocity from well logs indicate values of 3000 to 4000 m/s (Sabrina Paolacci, 2006, personal communication). Based on an extensive seismic refraction study of perialpine lakes, Finckh et al. (1984) obtained velocities of 1500 to 1900 m/s for the upper stratified sediments, 1900 to 2700 m/s for the lower glacial units and 3600 to 6700 m/s for the bedrock. For Lake Geneva, Vernet and Horn (1971) reported a velocity of 1600 m/s for lacustrine and glacio-lacustrine sediments and 2000 m/s for unstratified glacial sediments. More recently, Scheidhauer (2003) obtained velocities of 1450 to 1600 m/s for the unconsolidated sediments and from 1600 to more than 3000 m/s for the consolidated sediments.
Chapter 12
Discussion

12.1 Origin of till units

“An outstanding problem (in glacial marine stratigraphy) is the genesis of thick acoustically unstratified glacigenic sediments that are stratigraphically overlain by glacial-marine sediments” stated Syvitski in 1991. On high-resolution seismic records, till is difficult to differentiate from ice-loaded glacio-lacustrine sediments (i.e., proglacial subaqueous sediments overridden by the glacier), ice-contact deposits such as grounding-line fans, or debris flow deposits. All these sediments are diamictons containing many point sources which reflect acoustic energy in a disorganised manner, leading to a chaotic to semi-transparent seismic facies regardless of their origin (Syvitski et al., 1997).

In Western Lake Geneva, this facies has been encountered in units U4, U5, U7, U9c and U11c. The geometry of the Nyon and Copper push-moraines suggests that the acoustically unstratified units U9c and U11c post-date these readvances and correspond therefore to recessional deposits, such as grounding-line fans or debris flows. U7 is directly under the glacio-lacustrine sediments and must also correspond to recessional deposits. In Lake Annecy, coarse proximal deposits also appear with a transparent facies and their proglacial origin is inferred from the transition to stratified deposits towards the lake borders, interpreted as distal underflow deposits (Beck et al., 2001).

However, the lithology and the sedimentary process of units U4 and U5 are unknown. For the lithology, the best source of information for these units is the Zübo core in Lake Zürich (Lister, 1984b). Underneath the stratified glacio-lacustrine sediments, this long core revealed several units of uncompacted, muddy diamicct, alternating with deformed glacio-lacustrine sediments (unit 4b in Table 12-1). On Lake Zürich seismic sections (Hsü & Kelts, 1970; Giovanoli et al., 1984), these two types of sediments present a chaotic seismic facies and cannot be differentiated, indicating that the glacio-lacustrine sediments are too deformed to produce a stratified seismic facies. It is therefore probable that units U4 and U5 correspond to similar muddy diamiccts, with possible intercalations of deformed glacio-lacustrine sediments.

The progressive infill of the glacial basin with more than 200 m of subglacial sediments (Fig. E-10) requires creation of accommodation space below the glacier. This space was obtained by progressive thinning of the floating glacier. The limits between these seismic units correspond to strong reflections on the lake borders, which become weaker towards the lake center. These reflective levels were probably produced by ice shearing and compaction of the underlying sediments during ice readvances, which also correspond to a thicker glacier. During these readvances, the ice was resting on the lake borders but was still supported by water in the lake center, explaining the lower reflectivity in this zone. From the above considerations, it can be concluded that units U4, U5 and U7 were deposited during a general deglaciation associated to small readvances.

If U4 and U5 are muddy diamiccts, these have probably a complex origin. Part of the sediments is probably undermelt tills (Gravenor et al., 1984) formed subglacially by melting of the ice base and release of englacial debris. If one takes into account the oscillations of the ice front linked to oscillation of the ice base elevation, another part of the sediments probably corresponds to deformation tills formed during the readvances. If the oscillation were of high amplitude and led to the deglaciation of the basin, reworked proglacial sediments must also be present.

12.2 Eskers

12.2.1 Eskers or gravel drumlins?

The rectilinear path of the Petit Lac gravel bodies and the apparent presence of a steep stoss-side on longitudinal profiles (Figs. D-9 and D-10) might suggest gravel drumlins rather than eskers. Stratified sediments found within drumlins are generally older sediments eroded during the drumlinization and have thus no direct relation with
the drumlin genesis. However, some drumlins contain stratified sediments deposited during the drumlinization process. Two main types of stratified drumlins are reported in the literature: glaciofluvial infilling of subglacial cavities and lee-side glaciofluvial deposition. The first type of gravel drumlins, already presented in section 2.2.1, is related to the fluvial infilling of subglacial cavities during catastrophic subglacial meltwater floods (Shaw, 1983; Shaw & Kvill, 1984; Sharpe, 1987; Rampton, 2000). The second is related to the lee-side deposition of glaciﬂuvial sediments in a lee-side water-filled subglacial cavity during drumlinization (Dardis & McCabe, 1983; Hanvey, 1989; McCabe & Dardis, 1989; Dardis & Hanvey, 1994). Because neither of these models would explain the downflow widening of the conduit and the transition from coarse to fine deposits, the drumlin hypothesis was discarded for Western Lake Geneva.

12.2.2 Esker types

Currently, the formation process of long eskers is a matter of debate. They may form either time-transgressively near the ice margin during ice retreat, many smaller eskers joining to form finally a single large landform, or synchronously in a single continuous subglacial conduit. The longitudinal continuity of sedimentary sequences in the Petit-Lac eskers show that eskers of at least 2.5 km length can form synchronously in a single conduit.

According to a recent morphological classification of eskers into five types (Brennand, 2000), the Petit-Lac esker are of type II, i.e., short and subparallel eskers terminating in standing water. Type II eskers usually present the following characteristics: downstream grain size decrease, horizontal or downslope longitudinal profile, association to grounding-line or subaqueous fan sedimentation either superimposed over the coarser esker core or extending from it. The rectilinear path of the eskers indicates a strong hydraulic gradient in the esker direction and indicates therefore a steep ice surface gradient near the ice margin.

Cross-sections through the Petit-Lac eskers, and especially through the western one, are also very similar to multi-crested beaded eskers cropping out in Sweden (Hebrand & Åmark, 1989, Fig. 12-1). These esker systems would form time-transgressively by the deposition of successive ice-marginal sediment sequence, called beads, reflecting the positions of the receding glacier front. Each bead is characterized by decreasing grain size and descending upper surfaces in a down-glacier direction. These strong similarities suggest that each Petit-Lac esker may correspond to a single bead formed during a given ice front position. Evolution of subglacial drainage after the formation of this first bead did not favour the formation of other beads upflow.

Shreve (1985) related the morphology of tunnel fill esker cross-section to hydraulic conditions beneath the glacier and in particular the melting and freezing rates of the tunnel walls. According to his classification, sharp crested eskers are attributed to zones of net melting where the channel migrates vertically into the overlying ice as the floor fills with sediment, while multiple crested eskers are attributed to zones of lower melting rate where the channel has a tendency to migrate laterally rather than vertically. In the Petit-Lac, the western esker downflow transition from sharp-crested to multiple crested does not support Shreve’s theory and is rather related to a decreasing ice pressure and/or flow velocity.

12.2.3 Esker internal stratigraphy

Despite the high vertical resolution of our seismic data, the internal seismic stratigraphy of esker bodies was limited to the delimitation of the main sedimentary sequences for the western esker. Internal seismic stratigraphy of eskers presents the following difficulties:

- Strong dips on the esker flanks are poorly imaged by seismic reflection, because the seismic energy is not reflected towards the hydrophones (Fig. D-8).
• Longitudinal profiles following approximately the longitudinal axis of eskers are often contaminated by out-of-plane reflections. The large parabolas appearing on the unmigrated profiles transverse to the eskers (Fig. 12-2) clearly show the importance of this contamination. Besides, slight shifts from the esker path during acquisition lead to complex intersections between the 3D sedimentary structure and the 2D plane of seismic profile.

• Erosion surfaces corresponding to the truncation of older sequences by new sequences prevent the following of seismic reflectors over an extended area. These erosion surfaces are visible where they truncate layered units, but may be invisible where the erosion surface is parallel to the stratification of where the eroded unit is unstratified.

• Some parts of the esker deposits present a transparent facies, probably corresponding to massive and unsorted sediments deposited as traction carpets.

12.2.1 Esker location

In glacial valleys, eskers are usually located at the bottom of the valley (e.g., Mullins et al., 1996). In the Petit-Lac, the western esker high position on the western border of the valley results very probably from a lower ice pressure on the valley borders. In the central part of the Petit-Lac, where eskers are best developed, esker paths do not follow the path of the glacial valley: the former are straight while the glacial valley is bended. This suggests that the meltwater path was driven by the ice pressure gradient rather than the topography.

The presence of the eskers near the ice margin and their absence upflow, typical of type II eskers, can have two explanations. First, the meltwater may have come from the ice surface through sink holes (crevasses and moulins) only present at the ice front due to glaciotectonic constraints or cold-ice conditions up-glacier (Nienow et al., 1998). A variant of this first idea is the model of Shreve (1972) where englacial channels are dipping downflow and form an arborescent system perpendicular to the equipotential surfaces. According to this model, eskers could have been englacial in the northern part of the Petit-Lac. A second explanation is that meltwater flow may have existed upflow at the base of the glacier but its higher energy led to erosion or no deposition. Deposition would have only occurred near the ice margin where the water slows down due to channel widening. Given the presence of numerous diffractions upflow in the stratigraphical level corresponding to the eskers, we favour the latter hypothesis, illustrated in Fig. 11.5.

In the northern part of the Petit-Lac, indices of meltwater circulation are also present in other stratigraphic units (U2, U3, U5). In these units, reflective facies near the valley borders indicate marginal circulation of meltwater in lens-shaped conduits (Fig. D-3).

Eskers have also been reported in the Grand-Lac, at the bottom of the sedimentary sequence (Dupuy, 2005). According to the stratigraphical relationship between the Grand-Lac and the Petit-Lac, the Grand-Lac glacial sequence is younger than the Petit-Lac sequence. The Grand-Lac eskers may therefore have formed later in the deglaciation history, when the ice margin was further upflow.

12.2.2 Soft-bed eskers

The Petit-Lac eskers do not lie on a hard substrate but on a soft bed (also called “deformable bed”). Compared to a hard substrate, a soft bed constituted of till or other sediments is more easily erodible, more porous and more deformable leading to particular conditions for subglacial meltwater circulation (Johnson, 2002).

Eskers on soft bed are rare and soft-bed drainage models remain mainly theoretical (Brennand, 2000). According to Walder and Fowler (1994), the subglacial drainage of ice sheets lying on a deforming bed would not consist of an esker network, but rather of many wide, shallow, probably braided “canals” distributed along the ice-sediment interface and/or porewater flow. This theory is supported by the preferential occurrence of traditional eskers (large-scale, sinuous, steep-sided, morphological features) on hard-rock terrains (Clark & Walder, 1994). Few field cases exist to support this theory, but a well-known example is...
a soft-bedded, U-shaped, 5m deep, subglacial meltwater channel reported in eastern Germany (Piotrowski, 1999).

However, many field studies have shown that eskers are present in soft bed areas (e.g., Hart, 1996; Benn & Evans, 1998; Brennand, 2000; Sjogren & Brennand, 2005), in contradiction with the theory of Walder and Fowler (1994). These soft-bed eskers are generally cut down into the underlying till, indicating that subglacial channels were within both the ice and the till. This is not the case of the Petit-Lac eskers.

The presence of well-developed eskers in the Petit-Lac provides an interesting example for the study of subglacial meltwater circulation. Esker formation indicates that the underlying soft bed was not pervasively deforming during esker formation.

12.2.3 Comparison with onshore eskers

Digital elevation models presented in the second part of this work have not revealed any esker on the lake border. The chances of esker preservation are greater when these terminate in a standing water body, because this low-energy environment prevents erosion by proglacial braided streams (Banerjee & McDonald, 1975). They may however exist onshore but be invisible in the current landscape, due to burying under glacio-lacustrine sediments.

Eskers were however recognized further away from the lake. The Bois-de-Chêne eskers (section 7.2.1) are short and sinuous and do not represent an esker delta system. In the Ballens glaciofluvial complex (section 7.2.2) the sedimentology and inferred water-ending conditions are more similar to the Petit-Lac eskers.

12.3 Nyon and Coppet glacial readvances

Detailed seismic imaging of the Coppet and Nyon readvances allowed a better understanding of their geometry and of the sedimentary and glacio-tectonic processes associated with them. The present section discusses the implications of these new elements for subaqueous push-moraine formation, till-tongue formation, and the regional glacial history.

12.3.1 Push-moraine

Push-moraines have been described in various places based on field observations, but the scarcity of seismic investigations and subaqueous examples give particular value to the present study.

The geometry of the Nyon push-moraines is very similar to the push-moraines described by van der Wateren (1995) in Spitsbergen and Germany. In this model (Fig. 12-3), an external zone of folded, faulted and thrusted proglacial material lies over a major décollement plane. The ice compression against the bulldozed proglacial sediments led to the thickening of the ice thickness at the front. This new ice thickness was compensated by the formation of the overdeepened basin behind the compressive zone. Behind the basin, a horizontal surface is perturbed in places by compressive structures. In the NE part of line ia4 (Fig. D-19), the peak in the middle of the horizontal reflection corresponds probably to one of these compressive structures. The Petit-Lac push-moraines present the same spatial succession of structures, differing only from the model of van der Wateren by the absence of thrust sheets. This results probably from the higher ductility of the Petit-Lac sediments, related to their saturation in water and their small grain size.

In the proglacial environment, pushing takes place most readily when sediments have accumulated against the glacier margin, so that the advancing glacier can readily transmit a large force through a large sediment mass (Boulton, 1986). In the Petit-Lac, the accumulation of proglacial material in front of the advancing glacier was favoured by the geometry of the lake, which gets shallower and narrower towards the south.

The role of permafrost has often been invoked to explain the transmission of the deformation over a great distance ahead of the glacier front (e.g., Boulton et al., 1999). In the present case, this hypothesis can be clearly rejected because the glacio-lacustrine environment precludes the permafrost formation.

Note that the tectonic structures of the Petit-Lac push-moraines are also visible in other environments imaged by seismic reflection: orogenic fold-and-thrust belts, subduction zone accretionary wedges, mass movements and slumps. Orogenic belts also present thrust faults propagation over a décollement plane with older thrusts carried onto younger thrusts in a “piggy back” style, erosion of the tectonized sediments and transport to a wedge-shaped foreland basin, and progressive incorporation of the foreland basin sediments into the fold and thrust belt. The seismic imaging of accretionary wedges in subduction zones (Bangs et al., 2006) reveals a geometry extremely similar to that of the Petit-Lac push moraines.

Push-moraine glacio-tectonic structures also strongly resemble the fold-and-thrust belts developed at the front of subaqueous mass movements (Schneffelmann et al., 2005). These structures, revealed by high-resolution seismic reflection, are attributed to gravity spreading during the mass-flow deposition. 3D seismic imaging of large submarine slumps also show similar structures, with a basal shear surface ramping upwards approaching the slump toe (Frey Martinez et al., 2005). For the Petit-Lac push moraines, the main tectonic agent is clearly the ice front compression, but gravity spreading may have contributed to the deformation as a secondary agent.

12.3.2 Comparison of the Coppet and Nyon readvances

The extension of the deformation zone (or fold-and-thrust belt) in front of the inferred glacier position is
greater for Coppet than for Nyon (Fig. E-15). During the Nyon readvance, the small amount of bulldozed sediments could indicate a recently re-grounded floating ice-sheet. The deep basin of the eastern part of Lake Geneva would be compatible with a floating ice-sheet and a grounding line located at the eastern extremity of the lake, where the glacier came out of the Alps. This ice-sheet may have crossed the Grand Lac basin and reached the narrower and shallower Petit Lac, where it became grounded again. The Nyon push-moraine could thus have formed just after the “re-grounding” of the ice front and thus pushed a relatively small amount of sediment.

12.3.3 Glacigenic debris-flows vs. till-tongue model

Moscariello et al. (1998) and Chapron (1999) interpreted the Lake Geneva till tongues (units 9b and 11b) as waterlain tills deposited below a neutrally buoyant ice sheet, according to the model proposed by King and Fader (1986) and King et al. (1991).

King and Fader (1986) gave the first definition and interpretation of till tongues, which were slightly reworked by King et al. (1991). In the latter publication, which synthetizes seismostratigraphic studies of the Scotian and mid-Norwegian shelves, till tongues are defined as “wedge-shaped, acoustically incoherent, lateral extensions of massive till deposits which intertongue with stratified, ice-proximal glaciomarine deposits”.

Based on the stratigraphical relationship between till tongues and glaciomarine beds, their initial model (King and Fader, 1986) proposes the formation by subglacial melting of debris-rich ice just behind the grounding line (i.e., the line separating the floating ice from the grounded ice). Accordingly, the wedge shape of each till tongue is interpreted as an advance-retreat cycle of the grounding line (Fig. 12-4). King et al. (1991) modified this initial model by recognizing the contribution of “till aprons” (i.e., proglacial flow tills) at the distal end of the tongues and the erosion at the base of the till tongue during ice readvance.

The term “till-tongue” is not very well adapted for the description of semi-transparent wedges intercalated between stratified sediments because it already determines the lithology and glacial origin of the sediments. Moreover, it is implicitly associated to the “till-tongue model” of these authors, which supposes a subglacial formation. However, this term is commonly used in the description of high-resolution seismic sections in subaqueous proglacial areas for down-ice thinning wedges of transparent sediment intercalated between stratified sediments.

In the case of Western Lake Geneva, the presence of push-moraines behind the till-tongues indicates that the ice front was located behind them rather than over them. These wedges were therefore deposited in a proglacial setting rather than in a subglacial one.

Numerous authors have interpreted till-tongue-like deposits as proglacial glaciogenic debris-flows originating from material pushed by the glacier (Stravers & Syvitski, 1991; Boulton, 1996; Stravers & Powell, 1997; Dahlgren et al., 1998; Dahlgren, 1999; Plassen et al., 2004). Some have proposed a special name, such as till flows (Boulton, 1996) or debris lobes (Plassen et al., 2004). Such deposits are generally lobe-shaped in plan-view, lenticular in transverse section and wedge-shaped in longitudinal section. The main argument for the interpretation of these deposits as debris-flows is the similarity with non-glaciogenic debris-flow lenses recognized on seismic sections from the continental margins (e.g., Stoker et al., 1991; Aksu & Hiscott, 1992; Mahgoub, 1998).

The new seismic data presented here clearly indicate that the Petit-Lac till-tongues are proglacial debris-flows originating from the push-moraines. This new model should replace the subglacial interpretation of Moscariello et al. (1998) and Chapron (1999), who adopted the “till-tongue model” of King and Fader (1986).

In order to explain the till tongues, existing models of the Petit Lac deglaciation have presented an ice shelf, i.e., a floating, down-ice thinning ice mass uncoupled from its bed between the grounding line and the calving front (Moscariello et al., 1998; Chapron, 1999). The presence...
Fig. 12-4 King et al. (1991) original till tongue model (from Stravers & Powell, 1997).

Fig. 12-5 Stravers and Powell (1997) till-tongue model, interpreting these as debris-flow lenses coming from the downstream flank of a subaqueous moraine formed by the pushing of proglacial sediments (from Stravers & Powell, 1997)
of push-moraines rather indicates a steep ice front in order to push the sediments, which is not present in ice shelves but in tidewater glaciers, i.e., water-terminating glaciers present a steep ice cliff at the grounding line due to calving (important release of icebergs). As the receding Rhone glacier was temperate rather than polar, the presence of a tidewater glacier is logical because ice at the pressure melting point has a low tensile strength leading to important calving (Powell, 1984). Tidewater glaciers are more common today than ice shelves, with numerous examples in Alaska, Chile or Svalbard, and this was apparently also the case during the Quaternary glaciations (Benn & Evans, 1998).

A second important implication of our new interpretation of the till-tongue refers to the maximum position reached by the ice front during each readvance. Instead of reaching the toe of the till-tongue, the glacier only reached the internal part of the till-tongue, some 2 km upflow.

12.3.4 Readvance cycles

Chapron (1999) reported another till-tongue off Thonon (unit 2d of his thesis) attributed to another readvance, but it was not recognized by the more resent seismic study of Dupuy (2005), making its existence doubtful. On the original sparker seismic lines of Chapron (survey SP_AP_96 in Table 10-1), evidences for a till-tongue are relatively weak. Instead of having an intercalation of semi-transparent material between two stratified units, the data present the opposite situation: a thin and short lens of deformed stratified sediments intercalated between two semi-transparent units. This stratigraphical relationship does not correspond to the till-tongue definition of King and Fader (1986) or King et al. (1991). In addition, the presence of stratified material within unstratified material is only recognizable on one line and absent on neighbouring lines. Finally, these deposits are not associated with a thrust moraine. Correlation of these stratified sediments with the lacustrine sediments of the Petit-Lac is impossible because stratified sediments are acoustically absent at the boundary between the Grand Lac and the Petit Lac. However, the stratified sediments in the Thonon area are deeper in the sedimentary record than those of Coppet/Nyon. It is therefore probable that the stratified sediments recognized by Chapron (1999) correspond to subglacial deposits older than the Coppet and Nyon readvances.

In the Petit-Lac, although two readvances are recorded, more readvances could have occurred. The Coppet and Nyon readvances are visible on seismic because they were the last ones to reach their respective positions. By doing so, they have probably crushed and erased the sedimentary and tectonic imprint of older readvances of smaller extent.

In order to readvance, a tidewater glacier must be grounded, because the low tensile strength of temperate ice prevents an advance by floating (Powell, 1984). In the Petit-Lac, the accumulation of proglacial material at the front of the glacier reduced lost by calving during ice advance, but the overdeepening behind the push moraine increased calving during retreat. This configuration led probably to slow advances and rapid retreats, as it occurs in present-day tidewater glaciers (Lonne & Syvitski, 1997). This is in agreement with the rapid transition from glacial (unstratified) to glacio-lacustrine (stratified) deposits and the absence of large grounding-line fans or morainal banks (Powell, 1981). In the Grand-Lac, Dupuy (2005) also concluded to a rapid ice retreat. As calving rate is proportional to water depth, the ice retreat from the Grand Lac retreat must have been faster than in the Petit-Lac.

The Petit-Lac gets narrower and shallower downstream. This configuration increased the amplitude of the ice front oscillations, because ice advances were helped by the narrowing valley cross-section, which diminished calving and helped its progression, whereas ice retreats led to a size increase of the cross-section and a higher calving rate.

The velocity of the ice front movements is difficult to estimate. In present-day fjord glaciers, advance rates range from tens of meters to 12 km/yr (Liestol, 1969), whereas retreat rates range from 0.01 km/yr (Lavrushin, 1968) to 5 km/yr (Powell, 1991).

12.3.5 Lake level during the Nyon readvance

During deglaciation of the Geneva area, the Rhone glacier receded while the lake level decreased (Moscarielo et al., 1998). In Western Lake Geneva, the thick accumulation of subglacial sediments indicates that the glacier was floating. Because of the relative density of ice and freshwater, 90% of the volume of the floating ice mass was below water surface. Past lake level can therefore be estimated from the elevation of the upper and lower surfaces of the floating glacier, through the following formula: \( \text{LakeLevel} = \text{IceBase} + 0.9 \times (\text{IceTop} - \text{IceBase}) \). For the Nyon readvance (unit 11), the ice upper surface can be estimated from the elevation of the onshore fronto-lateral moraines (~450 m), and the ice base from the seismic stratigraphy (260 m in figure 11-2, assuming a velocity of 1450 for water and 1600 for the sediments). Based on the above formula, these two elevations give a lake level of 431 m, which corresponds to a previously reported 430 m lake level in the Lake Geneva basin (Moscarielo et al., 1998). This lake level was previously correlated with an ice front position between Laconnex and Geneva (Moscarielo et al., 1998). Geomorphological and seismic evidences presented here suggest that the 430 m lake level could rather be contemporaneous with the Nyon stage.

12.3.6 Datations

Unfortunately, no core reached the stratigraphic levels of the Nyon and Coppet readvances. According
Correlation with global cooling events during the deglaciation or with other readvance cycles in other areas is impossible because these two readvances only pertain to a segment of the Rhone glacier retreat and many other readvances probably occurred down- and up-flow. Besides, the glacier front position does not depend only on temperature but also on climate, basin geometry, glacier geometry, calving rate and substrate beneath glacier (Lonne & Syvitski, 1997). Even under the same conditions, two glaciers may have different behaviours. In the College Fjord (Alaska), two parallel tidewater glaciers originating from the same snowfield have shown opposite behaviours since 1931. The Harvard glacier has advanced at an average rate of nearly 20 m/y, whereas the adjacent Yale Glacier has retreated at a rate of approximately 50 m/y. This supports the hypothesis that their terminus behavior is largely the result of dynamic controls rather than changes in climate (Sturm et al., 1991).

12.4 Relation with surrounding stratigraphy

12.4.1 Relation with the Grand Lac seismic-stratigraphy

The seismic stratigraphy of the Grand-Lac is subdivided into four units: two glacial units and two lacustrine units (Dupuy, 2005). Like in the Petit-Lac, the Grand-Lac Quaternary sediments are affected by several erosion surfaces attributed to readvances of the Rhone glacier. The main erosion surface separates the two glacial units and corresponds to the base of the Nyon readvance in the northern part of the Petit-Lac. The lower glacial unit (glaciaire 1) corresponds to deposits older than the Nyon readvance, i.e., almost all the Quaternary infill of the Petit-Lac, whereas the upper glacial unit is interpreted as sediments deposited during ice retreat. This unit is only present in the westernmost part of the Grand-Lac and its thickness rapidly decreases towards the east. Most of the Grand-Lac Quaternary infill is therefore younger than the Petit-Lac infill.

The thickness of glacial sediments is much more important in the Petit Lac than in the Grand Lac. Dupuy (2005) suggested that an elevated Molasse zone at the western end of the Grand Lac may have protected the Petit Lac sediments from erosion.

12.4.2 Relation with the Geneva bay cores

The presence of gas blanking in the westernmost part of the lake (Fig. E-1) makes difficult the correlation of the seismic stratigraphy with the boreholes located in the Geneva bay, described and interpreted by Moscariello et al. (1998). However, a tentative correlation with the F4 core is proposed in figure 11-2.

- The overcompacted diamict B is bounded at the top by a major unconformity and could correspond to the seismic unit U1, attributed to an ancient glacial cycle.
- The gravels of unit C1 contain numerous wood fragments, one of which dated at 22'450±210 cal yr BP, and are covered by the till unit C2 in cores F2P and F4. They are interpreted as peri-glacial glacio-fluvial sediments deposited before the last readvance, between the deposition of seismic units U5 and U7. The datation could be older than the age of deposition of the sediments, because a wood fragment in the same unit was dated at 37'360±560 cal yr BP, clearly indicating sediment reworking.
- The till unit C2 and the overlying perturbed glacio-lacustrine deposits of unit D1 are interpreted as an advance retreat cycle and are attributed to seismic unit U7. This readvance could correspond to the Geneva city stage, during which the proglacial delta of Saint-Antoine was formed.
- Glacio-lacustrine units D2, D3 and E are correlated to seismic units U8, U10 and U12 to U14.

Correlation with the onshore stratigraphy around the Petit-Lac (section 9.3) is difficult because of the rapid lateral variations of facies and the pinch-out of the seismic units towards the lake borders. The interglacial proglacial gravel deposits of the Alluvion Ancienne seem absent in the Petit-Lac, in agreement with the Geneva boreholes located in the southern prolongation of the lake (Figs. 4-4 and 4-6). The Alluvion Ancienne was probably deposited in the Petit-Lac and later eroded by the following glacial activity. The thick deglaciation sequence of Lake Geneva (U2 to U12) correspond stratigraphically to the complex and discontinuous succession of till and glaciolacustrine units found on top of the Alluvion Ancienne in the Geneva area.

12.1 Comparison with other fjords of fjord-lakes

The sedimentary sequence of Western Lake Geneva is similar to that of other fjord-lakes in the perialpine area – e.g., lakes Zürich (Hsü et al., 1984; Lister, 1984b), Annecy (Van Rensbergen et al., 1998) and Le Bourget (van Rensbergen et al., 1999) – or in North America – e.g.,
Finger lakes (Mullins et al., 1996), Okanagan Lake (Eyles et al., 1991) and Shuswap Lake (Eyles & Mullins, 1997).

All these fjord-lakes present an elongated shape with an important overdeepening of the bedrock, reaching generally elevations below sea level. Their thick sedimentary infill was deposited almost exclusively during the last deglaciation, with no or little sediments older than the last glacial cycle. Unstratified or poorly stratified glacial sediments are overlain by stratified glacio-lacustrine to lacustrine deposits. Like in the Petit-Lac, the glacial sequence starts generally with a reflective semi-chaotic unit corresponding to basal meltwater circulation.

In all these lakes, the seismic stratigraphy focused on the upper part of the seismic record, with a fine subdivision of the upper glacio-lacustrine to lacustrine sediments but a coarser subdivision of the underlying glacial deposits. The finer subdivision of the glacial record achieved in the Petit-Lac results from several elements. The glacial sequence is exceptionally thick (more than 200 m) and the presence of compaction levels permits its subdivision. Besides, the presence of the Grand-Lac basin upflow acted as a sediment trap during the late deglaciation, resulting in a relatively thin cover of stratified sediments (maximum: 25 m) on top of the glacial sediments. Finally, the better penetration of the 5 and 1-inch³ airguns and the improved signal/noise ratio of the multi-channel recording achieved a much better imaging of the deeper sediments.

Among the perialpine lakes investigated by seismic reflection, Lake Zürich presents the greatest similarity with the Petit-Lac. Lake Zürich is divided into a northern upper lake and a southern lower lake. The upper lake acted as a sediment trap and limited the fluvio-glacial input sediment input into the lower lake. The same situation occurred in Lake Geneva where sedimentation decantation in the eastern part (Grand Lac) also prevented the late- to post-glacial infill of the western part (Petit Lac). However, the “trap” effect was more efficient in Lake Geneva, which is much larger than Lake Zürich upper lake.

In 1980, a long core named Zübo was drilled in the deepest part of lake Zürich (Hsü et al., 1984; Lister, 1984b), allowing the study of the entire sediment sequence (154.4 m) infilling the glacial valley cut into the Molasse (table 1). This sequence is very similar to the Petit-Lac sedimentary sequence from U2 to U14, and was also interpreted as a deglaciation sequence (Table 12-1). A large part of the infill consists in muddy diamicts and deformed glacio-lacustrine sediments, also found in the northern part of Lake Zürich (Schindler, 1974).

<table>
<thead>
<tr>
<th>depth (m)</th>
<th>litho-stratigraphic units</th>
<th>seismic stratigraphic units</th>
<th>lithology</th>
<th>seismic facies</th>
<th>Equivalent seismic units in the Petit-Lac</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-30.4</td>
<td>1-6</td>
<td>1-3</td>
<td>stratified sediments</td>
<td>stratified</td>
<td>U8, U10, U12-U14</td>
</tr>
<tr>
<td>30.4-80.6</td>
<td>7-10</td>
<td>4a</td>
<td>deformed glacio-lacustrine deposits with dropstones</td>
<td>sub-transparent with diffuse horizontal reflections in places. Some diffractions.</td>
<td>U3-US, U7, U9, U11</td>
</tr>
<tr>
<td>80.6-137.6</td>
<td>11-19</td>
<td>4b</td>
<td>alternation of deformed glacio-lacustrine stratified muds and compacted mud tills</td>
<td>low reflectivity. transparent to chaotic.</td>
<td>U2</td>
</tr>
<tr>
<td>137.6-154.4</td>
<td>20</td>
<td>4c</td>
<td>unlithified sand/gravels</td>
<td>disconinous reflections of medium amplitude.</td>
<td>U0</td>
</tr>
<tr>
<td>154.4-</td>
<td>Molasse bedrock</td>
<td></td>
<td>high amplitude reflection</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 12-1  Lake Zürich lithostratigraphy (Lister, 1984b) and seismic stratigraphy (Giovanoli et al., 1984), and tentative correlation with the Petit-Lac seismic stratigraphy.
Seismic interpretation of the Petit-Lac subglacial deposits has presented several challenges. The geometry of the deposits is complex, with numerous intersecting erosion surfaces and rapid lateral variations involving facies changes or vanishing of strong reflections. This complexity results from the highly dynamic glacial sedimentary system and particularly from the interaction between sedimentary processes related to ice and water flow. The dense grid of high-resolution seismic lines proved to be very adapted to this complex stratigraphy, highlighting its geometry and the processes of formation. Because of the complexity of the sedimentary record, numerous iterations were necessary to obtain a coherent stratigraphic model taking into account all the local constraints. Despite the willingness to propose a stratigraphy as reliable and consistent as possible, the model proposed here remains interpretative and subject to improvements by future researchers.

The Petit-Lac is located over a glacially-eroded bedrock valley narrowing and shallowing towards the southwest. This valley is filled with a thick sequence (up to 220 m) of glacial, glacio-lacustrine and lacustrine sediments divided into 14 seismic units (Fig. 11-2). Except for U1, the sedimentary sequence records the deglaciation of the last glacial cycle in a fjord-like environment occupied by a tidewater glacier with a steep, calving ice front. This general retreat was punctuated by readvances forming erosion surfaces below the glacier and push-moraines in front of it.

U1 represents remnants of glacial deposits older than the last glacial cycle, preserved in the deepest part of the lake and in secondary bedrock valleys. U2 represents gravel and sands deposited subglacially by meltwater circulation at the bottom of the glacial valley. U3 is a thick stratified units marking the beginning of the deglaciation, when the Rhone glacier became thinner and buoyant and allowed the formation of a subglacial lake. Subsequent glacial units (U4, U5, U7, U9, U11) are acoustically chaotic sediments deposited subglacially under the water table, while the glacier was thinning. These glacial units are bounded by synform erosion surfaces corresponding to small readvances of the glacier. These erosion surfaces are well defined in the northern part of the Petit-Lac and merge towards the south, where younger erosion surfaces truncate older ones.

The transition from the glacial to the glacio-lacustrine environment was progressive and started with the apparition of a marginal esker-fan system (U6). Esker formation was followed by a small advance-retract cycle leading to the deposition of U7. Then the ice front receded and stratified sediments were deposited in a glacio-lacustrine environment (U8, U10, U12). This retreat was punctuated by two readvances – Coppet (U9) and Nyon (U11) – producing large push moraines and proglacial debris flows. Finally, a lacustrine environment took place.

The presence of push-moraines clearly indicates a steep ice front and oscillations of the glacier front during the deglaciation. This is in contradiction with the deglaciation model of Lister (1984a) for Lake Zürich, which involved an ice body breaking up into blocks and icebergs.

Unstratified deposits intercalated between stratified glacio-lacustrine deposits, previously misinterpreted as subglacial till-tongues by Moscariello et al. (1998), are now interpreted as proglacial glacigenic debris flows due to slope instability on the push-moraine. This leads to a reevaluation of the positions for the Rhone glacier readvances, shifted by more than 2 km with respect to the previous inferred position.

As far as improvements in the seismic method are concerned, the processing flow inherited from former workers at the University of Geneva was “translated” to Seismic Unix for a faster, easier and more customizable workflow. Besides, two new steps were added to the processing flow: velocity analysis, which improved significantly the vertical sharpness of the seismic profiles, and spherical divergence compensation, which allowed the unbiased comparison of shallow and deep reflections. Finally, a method was developed to build a pseudo-3D seismic cube from a set of evenly spaced 2D seismic lines. This cube was of great help during the interpretation,
as it established a geometrical relationship from line to line and allowed viewing of the data following different orientations (inlines, crosslines, time-slices).

DEM analysis onshore and lacustrine seismic stratigraphy offshore proved to be complementary for the study of perialpine subglacial processes at the scale of Switzerland. Their association showed that the Petit-Lac eskers remain in the Petit-Lac glacial valley and do not extend onto the shores. With each method, much remains to be done to increase our understanding of glacial processes. Regarding DEM analysis, LiDAR data will get more easily available in the future and will permit the recognition of new glacial features and the accurate characterization of the landscape. Detailed mapping of fronto-lateral moraines could permit to correlate them from one glacial valley to another, based on pattern recognition in their distribution.

Regarding the future investigation of the Petit-Lac deposits, an increase in the density of the seismic grid would enhance the seismic stratigraphy, but the spatial resolution of a 2D seismic grid is always limited by the spacing of the 2D profiles (Cartwright & Huuse, 2005). The first consequence of this line spacing is spatial aliasing in the cross-line direction (i.e., the direction perpendicular to the 2D profiles). The 2D grid cannot resolve structures or sedimentary units smaller than the line spacing (e.g., 126 m in the case of Lake Geneva pseudo-3D survey). A second consequence of the 2D line spacing is related to the migration process, which assumes that all the recorded reflections come from a vertical plane located below the 2D lines. This assumption is only valid in the case of seismic lines exactly perpendicular to the geological structures. In the case of complex structures, numerous out-of-plane reflections (sidewipe events) are not corrected by migration.

Because of these limitations, a complete understanding of the complex Petit-Lac geometry would require a 3D seismic survey, like those already done in the Grand-Lac by Scheidhauer (2003) and Dupuy (2005). An extensive 3D seismic survey over the Petit-Lac would permit the reconstruction of buried surfaces through seismic geomorphology (Posamentier, 2004). Subsurface geomorphology could then be directly compared to the present-day morphology. Because high-resolution 3D surveys can hardly cover the entire surface of the Petit-Lac, we suggest to focus them on the eskers or the push-moraines. 3D investigation of the latter would reveal with accuracy the fault network linked to the glacial readvances.

Finally, a deep core through the entire Petit-Lac sedimentary sequence would be necessary to calibrate the seismic interpretation. The present study would provide the information necessary to choose a good location.
Part III

Appendices
Appendix A

GIS Methodology

A.1 Map digitization

Several maps had to be digitized during this work. Given the increasing demand for digital conversion of paper maps, a complete digitization workflow using Esri ArcGIS 9 is explained below in details for the bedrock map of Pugin (1988).

The bedrock map was scanned in several parts. In ArcMap, these parts were georeferenced and merged together. The map information was digitized in four shapefiles:
- bedrock_elevation.shp stores the bedrock contour lines with two attributes: bedrock elevation in meters a.s.l and sureness attribute (y for sure contour lines - continuous on the original map - or n for probable contour lines - dashed on the original map);
- bedrock_boreholes.shp contains the borehole with two attributes: elevation in meters a.s.l. and bedrock (y or n). If the borehole reaches the bedrock, the bedrock attribute is y and the height corresponds to the bedrock height. In the other case, the bedrock attribute is n and the height corresponds to the borehole deepest point;
- bedrock_seismiclines.shp stores the emplacement of the seismic lines;
- bedrock_questionmarks.shp contains the “question marks” of the original map in uncertain areas.

The contour lines digitization was done with ArcScan, a powerful raster-to-vector extension in ArcMap, offering the possibility to track semi-automatically dashed contour lines.

• In ArcMap, turn on the ArcScan extension (Tools >> Extensions >> ArcScan) and display the ArcScan toolbar (View >> Toolbars >> ArcScan).
• Load the raster to digitize and geo-reference it if needed.
• ArcScan needs a 1-bit (black and white) display. Right-click the raster layer and select Properties >> Symbology >> Classified. Classify your raster in 2 classes: black for the contour lines and white for the background.
• Create in ArcCatalog a polyline shapefile to store the contour lines (e.g., bedrock_elevation.shp). Add it to your ArcMap document. Start an “edit” session for this shapefile.
• Set the vectorization settings to suit your needs (ArcScan toolbar >> Vectorization >> Vectorization Settings). Several trials are necessary in order to find the appropriate parameters. For the bedrock map, we used the following ones: Intersection solution: none; Maximum line width: 6; Compression tolerance: 0.025; Smoothing Weight: 2; Gap closure tolerance: 12 (for dashed lines), no (for continuous lines); Fan angle: 35; Hole size: 1.
Appendix A - GIS Methodology

- Start digitizing using the Vectorization trace tool. When automatic tracking fails, press the S key while you click to enter points manually. If the raster line is closed (e.g., a circle), start at a point where the automatic tracking is likely to stop (e.g., at an intersection with another line) to prevent the tracking to pass again over its own path. Press F2 to finish the polyline.
- In the attribute table, enter the height of the new contour line in the Id column.
- Save your work regularly (Editor toolbar >> Editor >> Save Edits). If needed, modify the vectorization settings to suit your needs.

Finally, the contour lines were converted to a grid with a cellsize of 50 m. This was done with the Topo to Raster tool (ArcToolbox >> Spatial Analyst >> Interpolation >> Topo to raster), giving the bedrock map polygon as a boundary input file to limit the grid calculation to the original map extent.

A.2 DEM data pre-processing

A.2.1 SRTM

Switzerland SRTM data was obtained as three separate datasets from www.swissgeo.com. As each dataset had a different NoData value (West: 131, Centre: 45, East: -59), we used ArcGIS reclassification tool (ArcToolbox >> Reclass >> Reclassify) to assign them a common NoData value. This tool outputs integer values with a precision of 1 m, sufficient for the rather low vertical resolution of SRTM data (1.6 m). The three datasets were then merged and reprojected to Swiss coordinates.

A.2.2 LiDAR

The canton Vaud Lidar DTM was provided by the canton Vaud Service de l’Information sur le Territoire (SIT) as tiles of 4375 x 3000 m (a 16th of a Swiss 1/25’000 map) in ASCII Grid format. These tiles were converted from the ASCII Grid format to the ESRI Grid format for subsequent processing with the ESRI products. We used the following ArcInfo command repeatedly for each tile in an AML script, with the FLOAT keyword to specify a floating point data output:

```
ASCIIGRID D:\dtmgrid116334 _1m.asc D:\l16334 FLOAT
```

In order to increase display and calculation velocities, a lower resolution dataset was produced by subaveraging the original 1m grid to a 5 m grid. To resample the data with a bicubic interpolation, we used the following ArcInfo.GRID command on each tile:

```
K:\VD\tile69 _5m = RESAMPLE (K:\VD\tile69,5,CUBIC)
```

Then, the Vaud tiles and the Geneva DEM were assembled into a single file with the ArcInfo command Mosaic(). As this command accepts a maximum of 49 input grids, tiles had to be assembled into medium-sized grids and then the medium-sized grids into a single large grid. The advantages of a single large grid versus a collection of smaller grids are the following: easier data processing and handling and faster drawing thanks to the pyramiding of lower resolution grids on top of the original data. The command syntax is as follow. In case of overlapping, the first grids appear on top of the remaining ones.
MediumGrid1 = mosaic(tile01, tile02, tile03,...)
MediumGrid2 = mosaic(tile50, tile51, tile52,...)
...
LargeGrid = mosaic( MediumGrid1, MediumGrid2, MediumGrid3,... )

A.3 DEM attributes calculation

This section explains in detail the methods used for the calculation of the DEM attributes.

A.3.1 Low-cut elevation (Gaussian filtering)

Low-cut elevation is obtained in two steps: creation of a smoothed (low-pass) elevation model and subtraction of this smoothed elevation model from the original model, leaving only the highest frequencies.

The Lidar 5m DEM was smoothed with a Gaussian low-cut filter with a standard deviation of 50 cells (250 m) (Fig. A-1.a). Compared to the more common moving-average filter, the Gaussian filter offers the advantages of removing effectively the high frequencies without ringing effect. In the frequency domain, the plot as a function of spatial frequency

Fig. A-1 a) Gaussian operator with a standard deviation of 50 cells (250 m), as that used for the smoothing of the LiDAR DEM (cellsize = 5m). b) Frequency spectrum of (a) as a function of frequency. c) Frequency spectrum of (a) as a function of the spatial wavelength.
is also Gaussian-shaped (Fig. A-1.b). The plot as a function of spatial wavelength (Fig. A-1.c) shows that the operator completely removes wavelengths from 0 to ca. 150 cells (750 m), and progressively preserves larger wavelength: only 20 % of 250 cells (1250 m) wavelengths, more than 90 % for those larger than 1000 cells (5000 m). Once the smoothed DEM subtracted from the original one, the resulting low-cut DEM will have an inverse frequency spectrum, erasing 80 % of spatial wavelengths larger than ca. 250 cells (1250 m).

Practically, the smoothing is obtained by convolving the DEM with a convolution filter or kernel, which consists in a square grid representing the spatial repartition and the values of the weights around the focused cell. For this task, we take advantage of the possibility to decompose the Gaussian filter into X and Y components. Instead of convolving the DEM with a 2D Gaussian kernel, the DEM is blurred with a 1D Gaussian kernel along the X axis and then along the Y axis. This reduces the number of calculations and therefore the computing time (Fisher et al., 2000). For a Gaussian filter of 300×300 cells, a 2D kernel requires 90'000 (300 * 300) multiplications per cell, while the decomposition in 1D X and Y kernels requires only 600 (300 + 300) multiplications.

The desired 1D Gaussian filters can be calculated in Excel or generated automatically in a Python script (see gaussian_filter_construction.xls and GaussKern.py on the attached CD-Rom), with the following 1D Gaussian formula where σ represents the Gaussian standard deviation and determines the Gaussian width:

\[
G(x) = \frac{1}{\sqrt{2\pi}\sigma} e^{-\frac{x^2}{2\sigma^2}}
\]

Theoretically, a Gaussian kernel should have an infinite extent, but in practice a kernel 6 times larger than the Gaussian standard deviation is sufficiently accurate. Following this rule, we used a kernel size of 300 cells for a Gaussian standard deviation of 50 cells. In the spreadsheet, column A contains the input values corresponding to the cell number and ranges from -150 to 150 with an increment of 1. Column B contains the result of the Gaussian formula for a standard deviation of 50 specified in cell G2. Column C contains the values of B divided by the mean of B, to normalize the Gaussian operator so that its mean value is 1. This last operation is necessary for the later use of the values in Spatial Analyst with the FocalMean function. Based on the values of column C, we created two text files corresponding to the X (horizontal) and Y (vertical) directions (see kernel_gaussian_stdev50_X.txt and kernel_gaussian_stdev50_Y.txt on the CD-Rom). These operations are automated in the Python script GaussKern.py, also available on the CD-rom.

Once the X and Y kernels created, the Gaussian high-pass filtering was done with ArcMap >> Spatial Analyst >> Raster Calculator in three operations: smoothing in the X direction, smoothing in the Y direction, subtraction of the smoothed DEM from the original one. Note that Raster Calculator requires a carriage return at the end of the last command line.

```python
gevd5_gs50X = FocalMean(gevd5, WEIGHT, kernel_gaussian_stdev50_X.txt, NODATA)
gevd5_gs50 = FocalMean(gevd5_gs50X, WEIGHT, kernel_gaussian_stdev50_Y.txt, NODATA)
gevd5_gs50hf = gevd5 - gevd5_gs50
```

To avoid intermediate files, these three operations can be grouped into a single command line:
A.3.2 Hillshading

Hillshading simulates an illumination of the ground based on three main variables: illumination azimuth, illumination elevation and vertical exaggeration. The method takes advantage of the human brain assumption that the illumination is from overhead and shadows occur underneath objects. To create this illusion, the synthetic illumination must come from the northern half of the cardinal circle. If not, the eye and brain still assume that light comes from above and the map appears inside-out. The hillshading tool is available in ArcToolbox >> Spatial Analyst Tools >> Surface >> Hillshading. We used a solar azimuth of 315° (NW illumination) or 45° (NE illumination), a solar elevation of 45° and a vertical exaggeration of 2 to enhance the visibility of low-relief landforms.

For SRTM data, the input raster is in a spherical coordinate system (decimal degrees). To compensate the difference in measure between the horizontal ground units and the elevation z units, an appropriate z factor needs to be specified for that latitude. For the latitude of Austria (~ 48.3°), the appropriate z factor is 0.0000135 (ESRI, 2005). To obtain a vertical exaggeration of two, a z factor of 0.000027 was used.

A.3.3 Profile curvature

The profile curvature has been generated in ArcGIS 9 with the curvature tool (ArcToolbox >> 3D Analyst Tools >> Raster Surface >> Curvature). Default parameters were used to generate the curvature (mandatory) and the profile curvature. The best display is obtained with a white to black color scale where black corresponds to the concave values. This color scale is stretched onto the values range by the standard deviation method, using a value of 1 or smaller if needed to enhance the contrast. This stretching method shows how much a feature’s attribute value varies from the mean, by calculating the mean value and the standard deviation (ESRI, 2005).

The profile curvature output is a raster containing floating point (float) values but the range of values is small (from ca -400 to 400), resulting in an unnecessarily large file. To save storage space, the float values should be converted to an integer (int) type with the formula below in ArcMap >> Spatial Analyst >> Raster Calculator. Multiplication by 100 limits the loss of precision inherent to the conversion from float to int.

\[ [pcrv\_int] = \text{int} (100 \times [pcrv\_float]) \]

A.3.4 Openness

Positive openness was generated in MicroDEM, a freely available DEM analysis software (Guth, 2005). The openness algorithm calculates the mean value of the angles between the zenith and the horizon in eight directions within the specified radial distance (Fig. 3-5a). The output is expressed in hundredth of degrees (e.g., 90 degrees is expressed as 9000) and displayed with a black to white scale. Negative openness (Fig. 3-5b) follows the same principle but the view is directed downward. The processing flow, including the two-way format conversion between ESRI Grid and ASCII Arc grid, consists in the following steps:
Appendix A - GIS Methodology

- In ArcToolbox, convert the Arc binary grid to an Arc ASCII grid (ArcToolbox >> Conversion Tools >> From Raster >> Raster to ASCII). Make sure that the extension of the resulting ASCII file is “.asc”. Place this file in the folder X:\mapdata\DEMs created, where X:\ is the drive chosen during the installation of MicroDEM.
- In MicroDEM:
  - Press the in/out button and choose import >> ascii DEM >> ASCII Arc GRID.
  - If you are working with Swiss Grid data, choose rectangular for the digitizing datum in the Header dialog box. Press OK and a DEM with the extension .DEM will be created in E:\mapdata\DEMs. Close the dialog box.
  - Open the newly created DEM file, launch the positive openness calculation (Calculate >> Derivative Grid >> Openness upward) and specify the diameter of the openness calculator. For the DHM25 DEM, we used a diameter of 1000 m. Note that the calculation may take several hours.
  - Once the calculation finished, go to File >> Save DEM and save the openness map as an Arc ASCII grid.
- In ArcToolbox, convert the Arc ASCII grid to an Arc binary Raster (ArcToolbox >> Conversion Tools >> To Raster >> ASCII to Raster).
Appendix B

Seismic Unix Installation under Windows

B.1 Introduction

Seismic Unix (SU) is a seismic processing and research environment developed at the Center for Wave Phenomena (CWP) at the Colorado School of Mines. The package is distributed freely, with full source code. SU should be installed on a Unix-compatible operating system (e.g., Linux) but installation on Microsoft Windows is possible if one has previously installed Cygwin/X, also freely available. Cygwin/X consists of two parts. The first part, Cygwin, is a port of the GNU tools to Win32 and provides a UNIX-like environment in Windows without graphical interface. The second, Cygwin/X, is a port of the X Window System to Win32 and allows the display of Cygwin programs in standard Windows windows (Hunt, 2000).

These instructions were tested successfully under Windows 98, 2000 and XP. The last successful installation following these instructions was done in October 2005 with the following software versions: Windows XP sp2, Cygwin/X setup 2.457.2.2 and Seismic Unix 37.

The two first sections of this appendix present the installation of Cygwin/X and SU. The third section provides information and tips on how to use SU.

B.2 Cygwin/X Installation

B.2.1 Packages installation

Go to [http://x.cygwin.com](http://x.cygwin.com) and start the cygwin/X setup by clicking on “Install Cygwin/X now”. Go through the first install steps keeping the default options. When asked for a download site, choose one located near your location (e.g., mirror.switch.ch in Switzerland). At the Select Packages screen, you have to select the Unix components to install. Leave the default selection and add the components listed in Table B-1. To select an entire category, click on Skip or Default until Install appears. For packages, click on skip until the version number appears. Since Cygwin setup program automatically selects dependencies, do not unselect any dependent packages.

When you have selected the appropriate packages, click next to continue the installation. If download fails, you can try again with a different server without having to reselect the components. When Install is complete, press “Finish”. Put a shortcut to C:\cygwin\usr\X11R6\bin\startxwin.bat on the desktop. You will click on this shortcut every time you want to start Cygwin/X.
Appendix B - Seismic Unix installation under Windows

B.2.2 Modify/Add Windows environment variables

To complete the installation, you have to modify/add Windows environment variables. Under Windows NT/2000/XP, Right-click My Computer and select Properties >> Advanced >> Environment Variables. Create a new variable called DISPLAY and give it the value "127.0.0.1:0.0". Select the Path variable and click the Edit button. Add the line below at the end of the variable value:

;C:\cygwin\bin;C:\cygwin\usr\X11R6\lib

Under Windows 95/98, it is necessary to manually edit the file AUTOEXEC.BAT in a text editor to add or change the value of an environment variable. To add the DISPLAY variable, add the following line:

```
set DISPLAY=127.0.0.1:0.0
```

To update the PATH variable, you have to add the following string the end of the PATH line:

```
;C:\cygwin\bin;C:\cygwin\usr\X11R6\lib
```

Reboot your computer to make these changes effective.

B.2.3 Swiss-French keyboard installation

1. Search the web for the file xmodmap.ch_fr. Put it in the folder C:\cygwin\etc\X11\
Appendix B - Seismic Unix installation under Windows

2. Edit the file `C:\cygwin\usr\X11R6\bin\startxwin.bat` and add the following line at the end of the file:

```bash
xmodmap /etc/X11/Xmodmap.ch _ fr
```

3. Double-click the Cygwin icon on the desktop to open a cygwin terminal. Enter the following command:

```bash
cp /etc/X11/Xmodmap.ch _ fr ~/.Xmodmap
```

B.2.4 Test the installation

On the desktop, double-click the shortcut to `C:\cygwin\usr\X11R6\bin\startxwin.bat`. If installation is correct, an $X$ icon should appear in the Windows System Tray and a Cygwin/X terminal should pop up. The $X$ icon means that an $X$ server is running and that Cygwin/X programs can use it to display any graphical information in a window.

If you have problems to connect to the $X$ server, make sure that the internet TCP/IP protocol is installed. Go to Control Panel > Network and if TCP/IP is not listed, choose Add > Protocol … Add >> Microsoft >> TCP/IP >> Ok.

On Windows 95/98, you may also have to change the initial memory allowed to the DOS environment. Open a DOS window, right-click at the top of the window, click Properties >> Memory, set the conventional memory initial environment to the maximum (4096) and reboot.

B.2.5 Add a Cygwin prompt to folder context menu

When you start Cygwin/X, you are always in the home folder (i.e., `C:\cygwin\home\username`). To work directly in the folder where you have your seismic data, you can add in the folder context menu a Cygwin here option to start a Cygwin terminal within this folder. To do so, save the following lines to a file with a .reg extension (e.g., `cygwin.reg`). Run the file by double-clicking on it to write these new commands in Windows registry. If necessary, change the path to `bash.exe` to correspond to your installation.

```reg
REGEDIT4
[HKEY_CURRENT_USER\Software\Classes\Directory\shell\BashHere]
@="&Cygwin here"
[HKEY_CURRENT_USER\Software\Classes\Directory\shell\BashHere\command]
@="c:\cygwin\bin\bash.exe --login -c "cd '%1' ; exec /bin/bash -rcfile ~/.bashrc""
[HKEY_CURRENT_USER\Software\Classes\Drive\shell\BashHere]
@="&Cygwin here"
[HKEY_CURRENT_USER\Software\Classes\Drive\shell\BashHere\command]
@="c:\cygwin\bin\bash.exe --login -c "cd '%1' ; exec /bin/bash -rcfile ~/.bashrc""
```

B.3 Seismic Unix installation

B.3.1 Download

Go to `http://www.cwp.mines.edu/cwpcodes/` and download the latest version of Seismic Unix. It is distributed as a compressed file called `cwp_su.all.XX.tar.gz` where XX
corresponds to the version number. Create the folder C:\cygwin\cwp and download the file into this emplacement.

B.3.2 Decompression

When the download is finished, right-click on the C:\cygwin\cwp folder and choose “Cygwin here” in the context menu to run a Cygwin terminal. In this terminal, type the two following commands to decompress the file (replace XX by the version number):

   zcat cwp.su.all.XX.tgz | tar -xvf

B.3.3 Environment variables

To add environment variables, open the file C:\cygwin\etc\profile in a text editor and add the following two lines:

   CWPROOT="/cwp"
   export CWPROOT

Then, find the line

   PATH=/usr/local/bin:/usr/bin:/bin:/usr/X11R6/bin:$PATH

and add “:/cwp/bin” (without quotes) at the end, in order to get:

   PATH="/usr/local/bin:/usr/bin:/bin:/usr/X11R6/bin:$PATH:/cwp/bin"

On Windows NT/2000/XP, right click My Computer and select Properties >> Advanced >> Environment Variables. In the System variable part, click New... to create a new variable. Name the new variable CWPROOT and give it the following value: “C:\cygwin\cwp” (without quotes). If you are on Windows 95/98, you have to edit AUTOEXEC.BAT in a text editor to add the line below (and then reboot):

   set CWPROOT= C:\cygwin\cwp

B.3.4 Compilation

To run Seismic Unix, you must first compile the source code. Edit the file C:\cygwin\cwp\src\Makefile.config in a text editor. Makefile.config permits to set the options for the installation on your PC. Read carefully the comments and modify the file to correspond to a PC Cygwin installation, which is similar to a Linux installation. If you need help, compare your makefile with the makefile provided on the CD-ROM at the end of the present work, even though it may be outdated for your current version of SU. Once you have done the necessary changes to the makefile, you can compile the source codes. Open a Cygwin terminal and change the current location via:

   cd /cwp/src

Then, type the following commands to compile:
Appendix B - Seismic Unix installation under Windows

make install to install the basic set of codes, answer y (yes) to any question
make xtinstall to install the X-toolkit codes, see note below
make finstall to install the Fortran codes
make mglinstall to install Mesa/Open GL codes, optional
make xmininstall to install X-Motif codes, optional
make sfinstall to install SFIO materials and SEGREAD

**note:** if you get a “Nothing to be done for `xtinstall`” message when trying to compile the xtinstall package, compile xtinstall with the following command lines:

```
cd $CWPROOT/src/
cd ./Xtcwp; make; cd ..
cd ./xplot; make; cd ..
```

If problems occur during compilation, you can recompile the codes via:

```
made remake
make xtremake
make fremake
make mglremake
make xmermake
make sfremake
```

To test the installation, start Cygwin/X by double-clicking `startxwin.bat`. In the terminal, enter: `"suplane | suxwigb &"`. A synthetic profile should appear (Fig. B-1).

**B.4 Seismic Unix beginner guide**

This section presents basic informations about SU in association to Cygwin/X, particularly the way of displaying SU profiles. For more complete documentation on SU, you can download a clear and detailed SU manual (Stockwell & Cohen, 2002) from the CWP/SU website. To get information about a specific SU module (e.g., `sufilter` for frequency filtering), just enter the
module name at the command prompt. To get a description of a header field, type `sukeyword` followed by the header field name (e.g., `sukeyword dt`).

When you enter commands within a Cygwin window, be careful of the case (lowercase or uppercase) because Cygwin is case-sensitive. Note also that paths in Cygwin are expressed with a slash (/) and not a backslash (\) and that the “root” of the Cygwin file system corresponds to `C:/cygwin/`. If you want to access locations outside this root folder, you have to use the word `cygdrive`, which corresponds to My Computer under Windows. The table below presents translation examples between Windows and Cygwin paths.

<table>
<thead>
<tr>
<th>Windows path</th>
<th>Cygwin path</th>
</tr>
</thead>
<tbody>
<tr>
<td>C:\Cygwin</td>
<td>/</td>
</tr>
<tr>
<td>D:\seismic\Data</td>
<td>/cygdrive/d/seismic/Data</td>
</tr>
<tr>
<td>C:\Cygwin\Home\user</td>
<td>~</td>
</tr>
</tbody>
</table>

### B.4.1 Basic Unix commands

- **ls** lists the current directory content.
- **cd** changes directory. Example: `cd /cygdrive/d/seismic/Data`
- **clear** clears the terminal screen.
- **exit** closes the terminal.
- **&** if put at the end of a command, runs the command in the background in order to use the same terminal to enter another command.
- **|** is the piping symbol and allows you to redirect the output of a first operation as an input for the following one. The symbol `\` also works.
- **cat** concatenates files, e.g., `cat input1.su input2.su > output.su`
- **sh proc.sh** executes the shell script `proc.sh`. A shell script is a text file containing a sequence of commands, useful for writing a seismic processing flow. It is equivalent to a batch script in MS-DOS.

### B.4.2 Special keys

- **up/down arrows** browse through the previously entered command lines.
- **right-click m.b.** copy the highlighted text or paste the clipboard content. To enable copy/paste in Windows XP, you have to turn on the ‘quick edit’ mode for the command shell: Right-click on the window title bar, choose Properties, select Quick Edit, click OK, right-click again on the title bar and choose “default”.
- **tab** completes the filename. Example: to open the file `a01_neuchatel`, just type `a01` and press `tab`, the filename will complete automatically if there is no other file in that folder starting with the same characters.
- **space** goes to the following screen when reading online help.
- **q** quits the online help.

### B.4.3 Display a SU seismic section

To start working on your seismic data, you must first get a Cygwin command prompt in your data directory. First, run the desktop shortcut to `startxwin.bat`. Then, close the Cygwin terminal window and open a new one using the `Cygwin here` command that you
added to the Windows context menu (section B.2.5): in Windows Explorer, right-click on
the directory (not the file!) containing your seismic data and select Cygwin here. A terminal
appears and you are ready to work with your data.

To display a profile with the amplitude expressed in greyscale, type in the following
line (remember that Cygwin is case-sensitive):

```
suximage < ProfileName.su perc=97 &
```

If you prefer a wiggle display, enter:

```
suxwigb < input.su perc=97 &
```

To avoid typing the entire filename, you can enter the first letters and press the Tab
key to auto-complete it. The perc parameter is optional and its default value is 100. By
lowering this value, you increase the contrast of the profile. To zoom-in, draw a rectangle
with the mouse while pressing the left button. To zoom-out, left-click once on the profile.
When viewing seismic lines with suximage, you can browse through RGB colorscales
pressing r (next one) or R (previous one). Seismic data have a large number of associated
parameters. In SU data, the values of these parameters are stored in the header fields of
the traces, called header words. If you want to label the traces according to a particular
header word, use suxwigb and the key parameter. The following example labels the traces
according to the field record number (header word fldr):

```
suxwigb < input.su perc=97 key=fldr &
```

The suximage and suxwigb commands can be combined to other commands with the
piping symbol “|”, as in the following examples:

Display the first 500 traces:
```
suwind min=0 count=1000 < input.su | suximage perc=97
```

Display a pre-stack SU file and sort the traces by channel to track the bad channels:
```
susort < input.su tracf fldr | suximage perc=97
```

Display the 500 first traces, sorted by channel:
```
suwind min=0 count=1000 < input.su | susort tracf fldr | suximage
perc=97
```
Appendix C
Seismic Processing details

C.1 Introduction

In this section, we describe in detail the processing flow presented in figure C-1, which consists mainly in UNIX shell scripts executing a sequence of Seismic Unix commands. A shell script is a text file containing a list of commands to be interpreted by the operating system (Cygwin). The extension of such files is usually “.sh”. Shell scripts can be written in any text editor and can be run by typing “. scriptname.sh” at the Cygwin command prompt.

The processing flow (Fig. C-1) is constituted of three main parts: proc1.sh, velocity analysis and proc2.sh, presented in the three following sub-chapters. The scripts are annotated on the left with numbers referring to explanations below. Finally, the additional, optional processing steps are explained.

C.2 Proc1.sh

This first shell script prepares the data for velocity analysis. It requires a SEGY file as input.

```
#!/bin/bash
set +v
# set -x
for file in line01 line02 line03 line04 line05
do
  echo applying proc1.sh to $file ...
  # geometry editing
  # 1st channel corresponds to smaller offset
  SUSHW key=tracf a=1 b=1 j=11 <"$file".su |
  SUSHW key=offset a=7 b=7 j=11 |
  SUSHW key=sx a=112 b=0 c=1 j=11 |
  SUSHW key=gx a=111 b=-1 c=1 j=11 |
  SUSHW key=CDP a=223 b=-1 c=2 j=11 |
  # trigger delay corection
  SUWIND tmin=0.01 |
  SUSHW key=delrt a=0 |
  # spherical divergence correction
  SUGAIN tpow=1.0 |
  # bandpass frequency filtering (pre-deconvolution)
  SUFILTER f=100,140,1200,1800 |
```

Appendix C - Seismic processing details

**Legend**
- Data
- Processing step (software, SU module, shell script)
- Additional process requiring user input
- Decision
- Steps automatized in a processing script

---

**Fig. C-1** Flow chart for the seismic processing.

```plaintext
10 # deconvolution
SUPEF minlag=0.01 maxlag=0.028 pnoise=0.01 \ 
   mix=10 mincorr=0.1 maxcorr=0.35 |
11 # bandpass frequency filtering (post-deconvolution)
SUFILTER f=120,170,1200,1800 |
12 # direct wave muting
SUMUTE key=tracf ntaper=40 xmute=1,11 tmute=0.01,0.07 mode=0 |
13 # bad channels removal
SUWIND key=tracf reject=1,3 |
14 # sorting by CMP
SUSORT CDP offset >"$file".sort.su
15 # constant velocity stack
SUNMO smute=1.5 vNMO=1500 <"$file".sort.su | 
   SUSTACK >"$file".st1500.su
16 done
```

1. Indicates to the system that this script is written in `bash`, a widely used Unix shell.
2. Turns command echoing off.
3. This line is commented (i.e., ineffective) because it is preceded by a `#`. If the `#` is removed, the script will be debugged during execution. This function is useful to localize errors.
4. The filenames to be processed are passed as variables through a for loop. For each filename listed after the words for file in, the code located between the words do and done is executed and the variable named $file stores the filename.

5. Text preceded by the command echo will be displayed at the command prompt during the script execution. $file is the variable passed by the for loop. At the loop first cycle, the variable file has the value line01, so the message "applying proc1.sh to line01 ..." will be displayed at the command prompt.

6. The geometry is set according to figure 10-3 by changing header words of each trace with the command SUSHW. Here the commands are piped with the symbol '|'. This means that the output of a command is the input of the next one. SUSHW takes four parameters: (a) value on first trace; (b) increment within group; (c) increment between each group; (j) number of elements in group. In this example, channel recording order starts from the boat (channel 1 = smaller offset; channel 11 = larger offset). However, the channel recording order depends on the seismograph (Strataview or Geode) and on the presence of a channel converter: channel 1 may also correspond to the hydrophone further away from the boat. In that case, the geometry would be set with the following commands:

```
SUSHW key=tracf a=11 b=-1 j=11 <"$file".su |
SUSHW key=offset a=77 b=-7 j=11 |
SUSHW key=sx a=112 b=0 c=1 j=11 |
SUSHW key=gx a=100 b=1 c=1 j=11 |
SUSHW key=CDP a=213 b=1 c=2 j=11
```

7. The trigger delay is corrected by removing the first 10 ms of each trace with the SUWIND command. Note that the $min parameter is expressed in seconds. This operation automatically updates the header word $delrt, which indicates the time gap between the explosion and the start of the recording. It must be reset to 0 ms with the SUSWH command.

8. The parameter $tpow=1 multiplies each sample by the time value, in seconds. This compensates for the energy loss due to spherical divergence.

9. Before deconvolution, data must be filtered.

10. The deconvolution parameters depend on the “bubble effect” delay. The latter must be determined previously by autocorrelation of the data, through the following command: "SUACOR ncout=201 sym=0 <input.su >output.su". Note that the bubble effect delay may slightly vary from profile to profile, according to the airgun depth and pressure. Deconvolution itself is done with the SUPEF command, which requires 5 parameters. $mincorr and $maxcorr indicate the start and end of the autocorrelation window, which should include a maximum of reflections and multiples (or bubble ghosts) and a minimum of noise. Therefore, $mincorr must be set after the first arrivals and/or the reflection-free portion corresponding to water depth and $maxcorr before the final part of the trace where ambient noise dominates. The autocorrelation window length ($maxcorr - $mincorr) must be at least 8 times the total autocorrelation lag, to avoid bias (Yilmaz, 1987, p. 109). $minlag and $maxlag are estimated from the autocorrelation analysis. $minlag corresponds to the “gap length” and should be measured from the beginning of the autocorrelation to the wavelet second zero crossing. It approximates the length of the system wavelet (Benz, 1999). $maxlag corresponds to the “operator length” and should include the period of the bubble effect or the multiple to be suppressed. $mincorr, $maxcorr, $minlag and $maxlag must be expressed in seconds. $noise means “percent of noise” and corresponds to noise added to the data prior to deconvolution to stabilize the deconvolution. For predictive deconvolution, a small amount of noise is sufficient (typically 0.1 to 1 percent). Note that the backslash symbol (\) is added before the carriage return to split this long command into two lines.

11. After deconvolution, the data is filtered again with a narrower frequency range than in step 9.

12. This command removes the direct wave, which comes directly from the seismic source to the hydrophones without reflection.

13. This completely removes the unwanted seismic traces. Since it creates gaps in the traces sequence, it must be performed after the geometry editing and the first arrival mute.

14. Sorts traces by CMP and offset to prepare them for the following operations: velocity analysis, NMO correction and stack.

15. Performs NMO correction with a constant velocity of 1500 m/s and stacks the data.

16. End of the for loop.

C.3  Velocity analysis

The script velan.sh, included in the sample scripts from the SU package, performs a semblance analysis for a velocity range at a given CDP interval along a seismic line. For a reliable velocity analysis, CDPs should present numerous sub-horizontal reflections located at various depths. As these conditions are not encountered on every point along the profile, the SU script velan.sh has been modified in a new script called velan2.sh, which allows the user to choose the CDPs most appropriate to velocity analysis. These CDPs are selected before the analysis on a preliminary section stacked with a constant velocity of 1500 m/s. For each line, the velocity analysis was done as follows:
• Plot the stack obtained from proc1.sh with a command line similar to the one below, where \( f2 \) indicates the CDP number of the first trace and \( mpicks \) the name of the file where the picks will be stored:

\[
\text{SUXIMAGE <line01.st1500.su perc=97 f2=213 mpicks=line01.CDP.picks}
\]

• Position the mouse cursor on the representative CMP and type \( s \) to pick them. When you’re finished, type \( q \). The picks are saved in a two columns file (time, CDP) named after the \( mpicks \) value.

• Run the script velan2.sh with two arguments: “filename of the CMP-gather” and “list of selected CDP”:

\[
. \text{velan2.sh line01.sort.su “227 312 420 554 646 708 775 890 965 1066”}
\]

• For each CDP, the script velan2.sh will display a semblance velocity panel (Fig. C-2). The blue zones indicate the velocities giving the best semblance at a given depth, i.e., the best stacking velocities. Pick these velocities from top to bottom by positioning the mouse cursor over the blue zones and pressing the \( s \) key. When you have finished to pick velocities for a CMP, press \( q \). If you’re unsatisfied of your picks, you can type \( n \) at the command prompt to redo the last CDP. Do not worry about the error messages.

• The picks are saved in a velocity file (*.vel). Open it in a text editor to remove the comma at the end of the first line.

To enhance the quality of the velocity semblance panels, it is a common practice to group adjacent CDPs into super-CDPs. The averaging of several CDPs enhances the signal of the continuous horizontal reflections and cancels the noise. Our tests have shown that
this method is not adapted to the rapid lateral variations common in glacigenic sediments: instead of increasing the amplitude of the reflections, the averaging cancels out the discontinuous and/or dipping reflections (Fig. C-3).

C.4 Proc2.sh

```bash
#!/bin/bash
set +v
# set -x
for file in line01 line02 line03 line04 line05
done
for file in line01 line02 line03 line04 line05
do
  # NMO correction based on velocity analysis results
  SUNMO <"$file".sort.su par="$file".sort.su.vel smute=1.5 >"$file".NMO.su
  # stack
  SUSTACK <"$file".NMO.su >"$file".st.su
  # phase-shift migration
  SUNIGPS <"$file".st.su vmig=1500 dx=3.5 ltaper=10 \n  ffil=120,170,1200,1800 >"$file".m1500.su
  # conversion to SEGY
  SEGYHDRS <"$file".st.su
  SEGYWRITE <"$file".st.su tape="$file".st.sgy endian=0
  SEGYWRITE <"$file".m1500.su tape="$file".m1500.sgy endian=0
  # temp files deletion
  rm "$file".st1500.su
  rm "$file".sort.su
  rm "$file".NMO.su
  rm "$file".st.su
  rm "$file".m1500.su
  rm binary
  rm header
done
```

1. For the NMO correction, the file containing the velocity analysis results (*.vel) is passed to SUNMO via the `par` parameter. NMO induces a stretch of far offset data, particularly at small time values. The maximum stretch accepted is set by the `smute` parameter.

2. The module SUSTACK stacks adjacent traces having the same CDP number. In the case of spiky data, SUDIVSTACK (diversity stacking) is used with the following command: "SUDIVSTACK key=CDP winlen=0.064" "$file".st.su". Note that diversity stacking is not appropriate if the data contain anomalous null or low-amplitude samples. As
SUDIVSTACK gives them much weight to compensate their low amplitude, they will contaminate the stacked traces, resulting in very low-amplitude traces. With our data, this problem occurs in tow cases: 1) when the original data contains lower amplitude shots due to acquisition problems (e.g., records without airgun shot) and 2) in shallow water areas. In this latter case, the shallow reflections coincide with the zone partially muted during NMO to prevent excessive stretching (“stretch mute”), resulting in a muting of the first reflections for far offsets (Figs. C-4 and C-5). Therefore, the choice of the stacking method depends on the line. If the data are not spiked, use SUSTACK, else, use SUDIVSTACK. If the spiked data also contain low-amplitude records, remove the spikes manually or completely remove the spiked channels before and use SUSTACK.

3. Migration key parameters are \(v_{\text{mig}}\) and \(d_x\) which represent respectively the mean sediment velocity (m/s) and the trace spacing (m). The \(\text{ltaper}\) parameter sets the number of traces to taper for the left and right edges of the section, to avoid migration artefacts. The migration requires much processing time. To limit as much as possible the calculation time, the frequencies to migrate can be specified with the \(\text{ffil}\) parameter. As data have already been filtered with the same parameters (see proc1.sh), the filtering performed during migration won’t have any effect over the migration quality.

4. For data loading into Kingdom Suite, the unmigrated and migrated data are converted from SU to SEGY with the module SEGYWRITE. Before, SEGY headers must be written with the module SEGYHDRS.

5. Temporary files are deleted with the command `rm`.

C.5 Missing shots replacements

The triggering program `chronos.exe` used during acquisition was written by A. Pugin for a 9600 baud GPS connection and had to be modified to comply with the new Garmin GPS baud rate of 4800. Apparently, the source code we modified was not the final version and still contained a bug: the modified version (`chronos48.exe`) hanged from time to time at irregular interval and has to be closed and restarted. Each one of these crashes led to a gap in the acquisition ranging from 15 seconds to 2 minutes, depending on the seismic operator watchfulness and rapidity. The last shot coordinates were not written and `chrono48s.exe` had to be rebooted to continue recording. To fill the gaps, the missing coordinates have been interpolated with a small C program `cooOK.exe` (Appendix F) and blank shots have been added with the scripts `blank.sh` and `blank.list.sh`. `cooOK.exe` performs the following operations on the coordinates files:

- Search for missing coordinates due to crash, by calculating the time interval between shots. If the time interval between two shots is greater than 10, a crash has occurred.
- The number of missing traces is calculated as \([\text{crash\_duration} / \text{normal\_shot\_interval}] - 1\)
Appendix C - Seismic processing details

- Interpolation of missing coordinates.
- Creation of an output file with 4 columns: CMP, x, y, status. Status indicates whether the coordinates are original or interpolated.
- Creation of a log file with two columns indicating the last file number before each crash and the number of interpolated coordinates. Note that this log file is never deleted, information is always appended to it.

The information available in the log file generated by cooOK.exe is then used to generate the same number of blank seg2 files to replace the missing shots and fill the gaps within the lines. For each crash, blank shots are created and given the name of the preceding seg2 file (before the crash) plus a suffix (e.g., 10111_1.dat, 10111_2.dat, etc.). When the original and new seg2 files are sorted alphanumerically, this naming convention puts them in the correct acquisition order. This copying/rename operation is performed by two associated bash scripts: blank.sh and blank.list.sh. Once the blank files have been created, the seg2 (*.dat) files for each seismic line can be merged into a single Seismic Unix file (*.su) with the program seg2su.exe.

C.6 Pseudo-3D cube construction

A 3D cube is defined by an inline/crossline orthogonal system where inlines correspond to the acquisition direction. For the pseudo-3D cube, the 37 2D surveys are considered as inlines, resulting in an inline spacing of 126.11 m, that is very large compared to the crossline spacing (3.5 m). Crossline numbers are assigned starting from a “crossline guide” perpendicular to the surveys (Fig. C-6). For each 2D survey, the trace at the intersection with the crossline guide has been attributed the crossline number 10000. Traces on both sides of the crossline 10000 have then been numbered by incrementing towards the SE (10001, 10002, 10003,…) and decrementing towards the NW (9999, 9998, 9997,…). Originally, the SU file format has been designed for 2D seismic data and therefore lacks header words to store the 3D survey inline/crossline numbers. This information has been stored in the unused header words gelev and selev respectively, which store usually the receiver and source elevation. In practical terms, the operation can be decomposed in the following steps:

- In order to tie your 2D surveys together, create a new 2D survey (crossline guide, for instance) in KS to indicate the future 3D survey crossline direction. Load any segy file (ideally, a blank segy file). Enter the coordinates to make the new survey cross all the 2D lines perpendicularly, approximately in the middle of each 2D line (Fig. C-6).
- Display the crossline guide section and note the shot points corresponding to each intersection between the guideline and the 2D surveys. The intersections are indicated above the profile.
- Note also the acquisition direction of each 2D line: normal or reverse, normal corresponding to the lines acquired in the same direction as the crossline numbering (from NW to SE).
- Export 2D files as SEGY from Kingdom Suite, so that each trace header contains the shot point number in the tracr header word.
- Modify the Unix shell script 3D.sh (Appendix F) to suit your needs and run it.
Once the cube created with Seismic Unix, it can be loaded into Kingdom Suite and the following information must be provided:

- First/last inline and crossline numbers (inline: 101-137, crossline: 9595-10360)
- Increment between traces (1)
- Inline spacing and crossline spacing (126.11 m and 3.5 m, respectively). The inline spacing is obtained by dividing the crossline length by the number of intervals between inlines: $4540 / (37 - 1) = 126.11$ m.
- Coordinates for any 3 points of the survey (e.g., 3 corners of the pseudo-3D area)
- Byte number of the header words where the inline and crossline numbers are stored. (41 and 45, respectively)

C.7 Sparker Coordinates

The sparker coordinates are expressed as WGS84 latitude and longitude arc seconds. They are stored in the header bytes 73-76 (longitude) and 77-80 (latitude), except for the lines 1,2,3 and 30, which lack coordinates. Coordinates are stored with an unsufficient precision. Hence, they do not vary continuously but change every 10 to 20 traces, drawing a stair step path. In order to obtain smooth tracklines expressed in swiss coordinates, coordinates have been 1) extracted from the header files, 2) converted from arc seconds to Swiss coordinates in Excel and 3) smoothed by a 5 points running average (operator 1-1-2-1-1) in a small C program written on purpose (*CooSmooth5.exe*, see source code in Appendix).

C.8 Loading of SEGY 2D profiles into Kingdom Suite

The importation of numerous 2D SEGY files into Kingdom Suite can be a tedious and time-consuming task involving countless mouse clicks. We describe below a method to
make the importation much faster. Note that in Kingdom Suite a 2D line is called a 2D survey.

1. In Kingdom Suite, go to (Surveys >> Survey Management >> Create) and create a new empty 2D survey for each new SEGY to import.

2. Then, go to Surveys >> Import segy >> Import multiple segy files to open the Bulk import 2D SEG-Y traces dialog box, which displays an importation table containing 6 columns. Only the two first columns are of interest for a basic importation: the first one should contain SEGY file path on your hard disk and the second one the 2D survey name, as entered in step 1. To avoid the tedious task of filling the importation table in Kingdom Suite, save this table as a text file to modify it directly in a text editor. To save the table, go to File >> Save at the top left of the dialog box.

3. You now have a text file with the extension .lpt. Open it in a text editor (e.g., notepad) and replace the paths and names of the 2D surveys to suit your needs. Be extremely careful to preserve the layout of the lpt file: columns must be aligned to the right. If, you want to import your data into a datatype other than Amplitude, replace the word Amplitude (top left) by another data type name. Save and close the file.

4. Return to Kingdom Suite, in the Bulk import 2D SEG-Y traces dialog box of step 2. Go to File >> Open to open the lpt file. If you edited correctly the lpt file, the adequate survey names and file paths should fill the table. If not, return to step 3 to check the lpt file layout.

5. Click on the Import button to start the bulk import operation.

If the 2D surveys need to be reloaded after a re-processing, the lpt file can be reused and only steps 4 and 5 are necessary. When re-loading a SEGY profile in KS smaller than the previous one (because of a different processing flow, for instance), the old line has to be deleted first and recreated. Otherwise, a band of old data will not be overwritten by the new profile and will remain to the right side of the new profile.
Appendix D
Seismic Sections

Fig. D-1 Location of the seismic lines presented in this study, with reference to the figure numbers.
Appendix D - Seismic sections

Fig. D-2  Gas blanking on the seismic line AG1_JF_05.09. Note the sharp boundary between the gas-rich and gas-free areas and the preservation of primary reflections in the deepest part of the record.

Fig. D-3  Uninterpreted and interpreted transverse seismic section AG1_JF_05.01, located in the northern part of the Petit-Lac. Vertical exaggeration: ~5 x.
Fig. D-4 Uninterpreted and interpreted transverse seismic section AG5_AP_96.ag7, located in the central part of the Petit-Lac. Vertical exaggeration: ~5 x.

Fig. D-5 Uninterpreted and interpreted transverse seismic section AG5_AP_96.ac12, located in the southern part of the Petit-Lac. Vertical exaggeration: ~5 x.
Fig. D-6 Two seismic profiles showing the Célyigny faults (F) affecting the Molasse bedrock.

Fig. D-7 Uninterpreted and interpreted transverse section AG5_AP_96.ag04, showing the eastern paleo-valley, off Anières.
Fig. D-8  a) Selection of transverse profiles through the pseudo-3D survey showing the downflow evolution of the eskers, from reflective to semi-transparent. In U6, reflections are drawn in facies 6a, whereas the reflection-free areas pertain to facies 6b. b) Location map of the profiles.
Fig. D-9 Uninterpreted and interpreted longitudinal line AG1_IF_03.c2a following approximately the axis of the western esker. Note the downflow thickening of the esker and facies change from reflective (U6a) to semi-transparent (U6b). Upflow, the gravel body turns gently to the west, moving progressively away from the seismic profile. As the profile moves away from the gravel body crest, it crosses a second crest, stratigraphically below the first one, giving the misleading impression of a drumlin stoss side. Vertical exaggeration: 5x.
**Fig. D-10** Uninterpreted and interpreted longitudinal line AG1_JF_03.e following the axis of the eastern esker. Note the sudden increase of the conduit height, leading to the transition from a reflective (6a) to a semi-transparent/chaotic facies (6b).

**Fig. D-11** Uninterpreted and interpreted longitudinal line AG1_JF_03.d revealing south-dipping reflections in U7, interpreted as foresets deposited during ice front retreat. Vertical exaggeration: ~5x.
Fig. D‑12 Uninterpreted and interpreted section AG1_JF_05.13, showing weak stratifications in the upper part of U7. Vertical exaggeration: ~5x.

Fig. D‑13 Zoom on the eastern part of seismic section AG1_JF_05.15 (uninterpreted and interpreted), showing the earlier onset of glacio-lacustrine sedimentation in this area.
Fig. D-14 Uninterpreted and interpreted composite longitudinal section (AG1_JF_05.11 and AG1_JF_04.27265) showing the Coppet readvance. Debris flows originate from the upper surface of the push moraine. As the glacier and the deformation front advance, the debris flows are progressively incorporated into the push moraine. Given the lack of stratifications in the debris flows, thrust faults are invisible on the seismic section, but they probably exist in the most proximal part of the debris flow (dotted line). The figure is the combination of two seismic profiles. Because they are not exactly aligned in plan view, they were shifted in time (3 ms) to align the seismic units vertically. Vertical exaggeration: ~10 x.
Appendix D - Seismic sections
Appendix D - Seismic sections

Fig. D-15 Uninterpreted and interpreted zoom on the fold and thrusts of the Coppet readvance, on section AG1_Mb_02. Vertical exaggeration: ~5 x.
Fig. D‑16  Uninterpreted and interpreted transverse airgun profile AG1_JF_04.1615 showing the Coppet readvance. The lacustrine deposits are slightly tectonized. No fault plane is visible because the compression is perpendicular to the profile direction. Flow is out of the page. m = lake bottom multiple. Vertical exaggeration: ~10x.

Fig. D‑17  Uninterpreted and interpreted transverse airgun profile AG1_JF_04.1314 showing the lenticular shape of the Coppet debris-flow wedge in cross-section and channels (ch) cut into stratified sediments and filled with acoustically semi-transparent material. Flow is out of the page. m = lake bottom multiple. Vertical exaggeration: ~10x.
Appendix D - Seismic sections

Fig. D-18 Uninterpreted and interpreted longitudinal seismic section SP_AP_96.16 showing the Coppet readvance and the reflective facies at the toe of the debris-flow wedge. a) original data, b) amplitude envelope, c) interpretation. Vertical exaggeration ~10 X.
Appendix D - Seismic sections

Fig. D-19  

(a) Interpreted 15-inch³ airgun profile acquired by David Dupuy (UNIL), showing the Nyon push-moraine. Due to the lower resolution of the seismic source, the folded glacio-lacustrine stratifications are not imaged, but their lower contact is very reflective. Note the overdeepened basin behind the push-moraine. The letters on top of the profile refer to the push-moraine model of van der Wateren (1995), presented in Fig. 12-3. Vertical exaggeration: 5x.

(b) Interpreted profile AG1_JF_03.c2e revealing the stratifications of the folded and thrusted glacio-lacustrine sediments. Vertical exaggeration: 10x.
Chaotic reflections due to peg-leg multiples in overlying unit.

Fig. D-20  a) Transverse profile SP_AP_96.17; b) Transverse sparker profile SP_AP_96.ip1, slightly more distal. The reverse fault orientation towards the valley sides indicates that the pushing was more important in the valley center.
Fig. D-21  Uninterpreted and interpreted seismic section AG1_JF_03.q showing two Rhone glacier readvances - Coppet and Nyon push moraines and their ‘till-tongues’ - lying on top of older till and/or glacio-lacustrine sediments. The apparent dips of the glacio-tectonic structures are lower than the true dip, because the seismic line is not exactly perpendicular to the glacier compression axis. Overlying top units are Late-Glacial and Holocene lacustrine sediment. Reflection numbers refer to Versoix seismic stratigraphy according to Girardclos et al. (2003, 2005), summarized in Fig. 11-8. Three mass slides, crossed perpendicularly, are named ‘S7’, ‘S6’ and ‘1F’. These are coming from the nearby lake shore and were triggered at different times; ‘m’ indicates the lake bottom multiple. Vertical exaggeration is about 35 times.
Appendix E

Seismic Maps
Fig. E-1  Map of areas affected by gas blanking in Western Lake Geneva.

Fig. E-2  Molasse bedrock isochron map (ms TWT), with location of the Céligny strike-slip fault.
Appendix E - Seismic maps

Fig. E-3 Isopach map of Quaternary sediments.

Fig. E-4 Isochron map of U2 upper surface.
Fig. E-5 Isochron map of U3 upper surface.

Fig. E-6 Isochron map of U4 upper surface.
Fig. E-8 Map of the eskers in the Petit-Lac. Darker tones indicate sharp-crested, high-reflectivity eskers, while lighter tones indicate flat, low-reflectivity eskers. In the eastern esker, the white star marks a sudden flow expansion and a loss of reflectivity.

Fig. E-7 Isochron map of U6 top.
Appendix E - Seismic maps

Fig. E-9 Isochron map of the glacio-lacustrine sediment base.

Fig. E-10 Isopach map of the glacial sediments.
Appendix E - Seismic maps

Fig. E-11 Isochron map of the Coppet readvance base.

Fig. E-12 Isopach map of the Coppet readvance
Appendix E - Seismic maps

Fig. E-13 Isochron map of the Nyon readvance base.

Fig. E-14 Isopach map of the Nyon readvance.
Fig. E-15 Interpretation map of the Coppet and Nyon glacial readvances.

Fig. E-16 Map of the slumps associated to the reflector 14. Arrows indicate the source.
Fig. E-17: Lake bottom isochrone map.

Fig. E-18: Isopach map of the glacio-lacustrine and lacustrine sediments, including till-tongues.
Appendix F

CD-ROM Content

The CD-ROM attached to this work contains mainly GIS data, animations through the 3D seismic cube, small softwares and processing scripts, all listed in the table below. All these files were produced during this study, except where specified.

Table F-1 CD-ROM contents.

<table>
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<tr>
<th>Directory/</th>
<th>File Name</th>
<th>Description</th>
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<td>GIS/data/</td>
<td>TatukGIS_Viewer_1_8_8_348.exe</td>
<td>installation program of the freeware TatukGIS viewer. This software must be installed to see the three projects below. Also opens any GIS data from this CD-ROM.</td>
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<td>bedrock_elevation.TTKGP</td>
<td>TatukGIS project with bedrock elevation and Quaternary thickness</td>
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<tr>
<td></td>
<td>bedrock_morphofacies.TTKGP</td>
<td>TatukGIS project with the Molasse lithology and geomorphological facies</td>
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<tr>
<td></td>
<td>glacial_landforms.TTKGP</td>
<td>TatukGIS project with the glacial landforms</td>
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<td>bedrock/</td>
<td>elevation of the bedrock in Western Switzerland (ESRI binary grid)</td>
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<td>thickness of Quaternary sediments in Western Switzerland (ESRI binary grid)</td>
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<td>_glacial_shp/</td>
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<td>shapefile (line) glacial ridges in the Dombes area (France)</td>
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<td>shapefile (polygon) main drumlin (and ribbed moraine) fields of the Alpine foreland</td>
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<td>drumlins.shp</td>
<td>shapefile (polygon) individual drumlins, mainly in the Swiss Alpine foreland</td>
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<td>shapefile (point) centre of mass of each drumlin</td>
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<td>shapefile (point) highest point of each drumlin</td>
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<td>shapefile (line) longest axis of each drumlin</td>
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<td>shapefile (polygon) outline of the till patches in the Joux valley area</td>
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<td>shapefile (point) key locations for this study</td>
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<td>molasse_MorphFacies.shp</td>
<td>shapefile (polygon) Western Switzerland Molasse: geomorphological facies</td>
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</table>
Appendix F - CD content

Molasse_streamlining_around_PtLac.shp  shapefile (line)  large Molasse lineaments around western lake Geneva
moraines.shp  shapefile (line)  fronto-lateral moraines on the Swiss plateau
RibbedMoraines.shp  shapefile (polygon)  individual ribbed moraines, mainly in the Swiss Alpine foreland
WGM_Alps_Jura.shp  shapefile (polygon)  extent of the Wurmian Glacial Maximum for the Alpine and Jura ice sheets

* srtm90/*  Extract of the SRTM-90 digital elevation model (CIAT, 2005) covering Switzerland and neighbouring areas (ESRI binary grid)

* GIS/tools/*

GaussFFT.py  Python script to compute and plot a Gaussian function and its Fourier transform.
GaussKern_1.py  Python script to generate X and Y kernels formatted according to the requirements of the FocalMean function of ESRI Spatial Analyst
GaussKern_2.py  Same as GaussKern_1.py, but processes directly an ESRI raster with a gaussian low-pass filter, using ESRI Spatial Analyst. Requires the Spatial Analyst extension
kernel_gaussian_stdev50_X.txt, kernel_gaussian_stdev50_Y.txt  Gaussian horizontal and vertical kernels for use in Spatial Analyst with the FocalMean function.

* Seismic/data/*

3D_inlines_envelope.mpeg  3D animation through the seismic cube envelope attribute (inlines)
3D_timeslices_envelope.mpeg  3D animation through the seismic cube envelope attribute (time slices)
lines_from_SW_to_NE.avi  animation through the seismic cube (inlines)
timeslices_envelope.avi  animation through the seismic cube envelope attribute (time slices)

* Seismic/tools/*

3D.sh  shell script converting a pseudo-3D constituted of several 2D lines into a single 3D cube (see section C.6)
blank.sh, blank.list.sh  creates blank seismic traces to fill in the gaps during the acquisition (see section C.5)
cooOK.c, cooOK_input.txt  C program interpolating missing coordinates in seismic acquisition gaps, with sample input file (see section C.4)
CooSmooth3.c, CooSmooth5.c, CooSmooth_input.txt  C programs smoothing the sparker coordinates with a mean filter of three or five values, with sample input file (see section C.7)
distance.sh  shell script calculating the total length of a set of 2D seismic lines
KS_loading.lpt  sample ascii text file used to load rapidly the 2D seismic profiles in Kingdom Suite (see section C.8)
proc.sh  shell script processing seismic data with a constant velocity (no velocity analysis)
proc1.sh, proc2.sh, proc.list.sh  shell scripts with the processing flow to be applied before/after velocity analysis, and sample list of files to be processed through proc1.sh and proc2.sh
Seg2su.exe  program to convert seismic data from SEG2 to SU format (author unknown)
su2Evf.c, evf2su.c  C programs to convert seismic data from SU to EVF format, and vice versa
su2Segy.sh, segy2su.sh  shell scripts to convert seismic data from SU to SEGY format, and vice versa
TrierWGS84.xls  spreadsheet performing two operations: interpolation of the shot coordinates between the recorded GPS coordinates and conversion of these new coordinates from the latitude/longitude system WGS84 to the Swiss grid CH1903. Modified from an anonymous pre-existing spreadsheet (WGS84.xls).
velan2.sh, velan2.list.sh  interactive script for velocity analysis to be performed between proc1.sh and proc2.sh (see section C.3 for details)
velmod.sh  This script interpolates the stacking velocities obtained from the CMP analysis into a continuous velocity model and then converts it to an interval velocity model. For the interpolation, we used the SU modules UNISAM and UNISAM2. UNISAM first converts 1-D picked velocities to a uniformly sampled velocity function. These 1-D velocity functions are then interpolated with UNISAM2 to create a 2D model. The model obtained gives the stacking velocities (RMS velocities) as a function of time and CDP. The model is then converted to an interval velocity model with the module VELCONV.

* PhD/*

PhD_Fiore_2007.pdf  Pdf version of the PhD thesis
References


Knavel, T., 1999, Late-glacial and Holocene vegetation history and dynamics as shown by pollen and plant macrofossil analyses in annually laminated sediments from Soppensee, central Switzerland. Vegetation History and Archaeobotany, 8(3), 165-184.


Whitman, D., 2002. Visualization and Analysis of Gridded LIDAR Digital Elevation Data with ArcView Spatial and 3-D Analyst, Department of Earth Sciences and International Hurricane Center, Florida International University.


