Chapter 2

Quaternary Glaciations in Austria

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2.1. INTRODUCTION

From the onset of rhythmic loess accumulation, at the turn from the Gauss to the Matuyama magnetic Chron, a frequent variation from humid warm to dry cool climate with loess and occasional gravel accumulation occurred until the end of the Matuyama Chron. Remnants of four glaciations (Günz, Mindel, Riss, Würm) within the Eastern Alps and their foreland have long since been known. More recently, evidence for an additional cold stage between the two older ones was found. As a result of a major cooling and build-up of piedmont glaciers in the foreland, the four glaciations show a complete succession of terminal moraines with glaciofluvial terraces connected to them. The last interglacial/glacial cycle can easily be reconstructed climatologically and by sediment development. It may serve as a model for understanding the climatic conditions of the older ones, which had very similar successions.

2.2. THE COURSE OF THE QUATERNARY

Around the Neogene/Quaternary boundary (Gibbard et al., 2009; Mascarelli, 2009), at 2.58 Ma BP, the drainage pattern of the Eastern Alps was generally fully developed. Loess accumulation began along the Danube north of the Eastern Alps at that time (Frank et al., 1997), at the Gauss/Matuyama Chron boundary (Fig. 2.1). According to palaeontological investigations, the Neogene to the Quaternary boundary period was characterised by a moderately warm, humid climate alternating with cool, dry periods. There were frequent repeated changes of climate, but lacking any reference to glacial events. Within the Eastern Alps, no sediments from this early phase of the Quaternary have been preserved, due to the high relief energy and the later glaciations. Sediments from this period, mostly loess, are found only in the Alpine foreland along the Danube and its tributaries.

The loess section at Krems shooting range (Fink et al., 1976) reveals a repeated change of dry-cold (*Pupilla* fauna), warm dry (*Striata* fauna) and warm humid (*Chilostoma* fauna) conditions at the beginning of the Quaternary (Fink et al., 1976; Frank and van Husen, 1995). The sequence contains evidence of 17 interglacials. They post-date the Olduvai Event and thus are younger than Marine Isotope Stage (MIS) 63. The loess was certainly deposited under dry, cold conditions. However, no evidence was found for any climatic deterioration strong enough to generate glaciation.

2.3. THE FOUR ALPINE GLACIATIONS

Following younger interpretations of the marine oxygen isotope record, MIS 24 at about 0.9 Ma represents a clear transition to a different style of climatic regime characterised by more extreme glaciations than the previous period (Shackleton, 1995). This may explain why within the Eastern Alps and their foreland in Austria no glacial deposits older than the Günz Glaciation, *sensu* Penck and Brückner (1909/11), have been found.

In the Salzach, Traun and Krems river areas, terminal moraines of this glaciation are connected to terrace bodies (Weinberger, 1955; Kohl, 1974). These *Ältere Deckenschotter sensu* Penck and Brückner (1909/11) form part of a widespread gravel cover that occurs between the rivers Traun and Enns. In terms of gravel composition and age, this unit is a genetically polymict body (van Husen, 1980, 1981). It probably accumulated over a long time spanning several cold stages. Reworking and lateral erosion by the rivers resulted in the incorporation of older deposits into what appears now as a single terrace accumulation.

In the Alpine foreland, east of the Salzach River (Fig. 2.2 and 2.3), remnants of three glaciations younger than the oldest represented above can be easily identified (Weinberger, 1955; Del Negro, 1969; Kohl, 1976; van Husen, 1977, 1996; Sperl, 1984). Knowledge regarding the extent, development of tills, terminal moraines, glaciofluvial sediments and weathering is good enough to allow the reconstruction of the individual ice streams and tongues of the piedmont glaciers (Penck and Brückner, 1909/11; Sperl, 1984).

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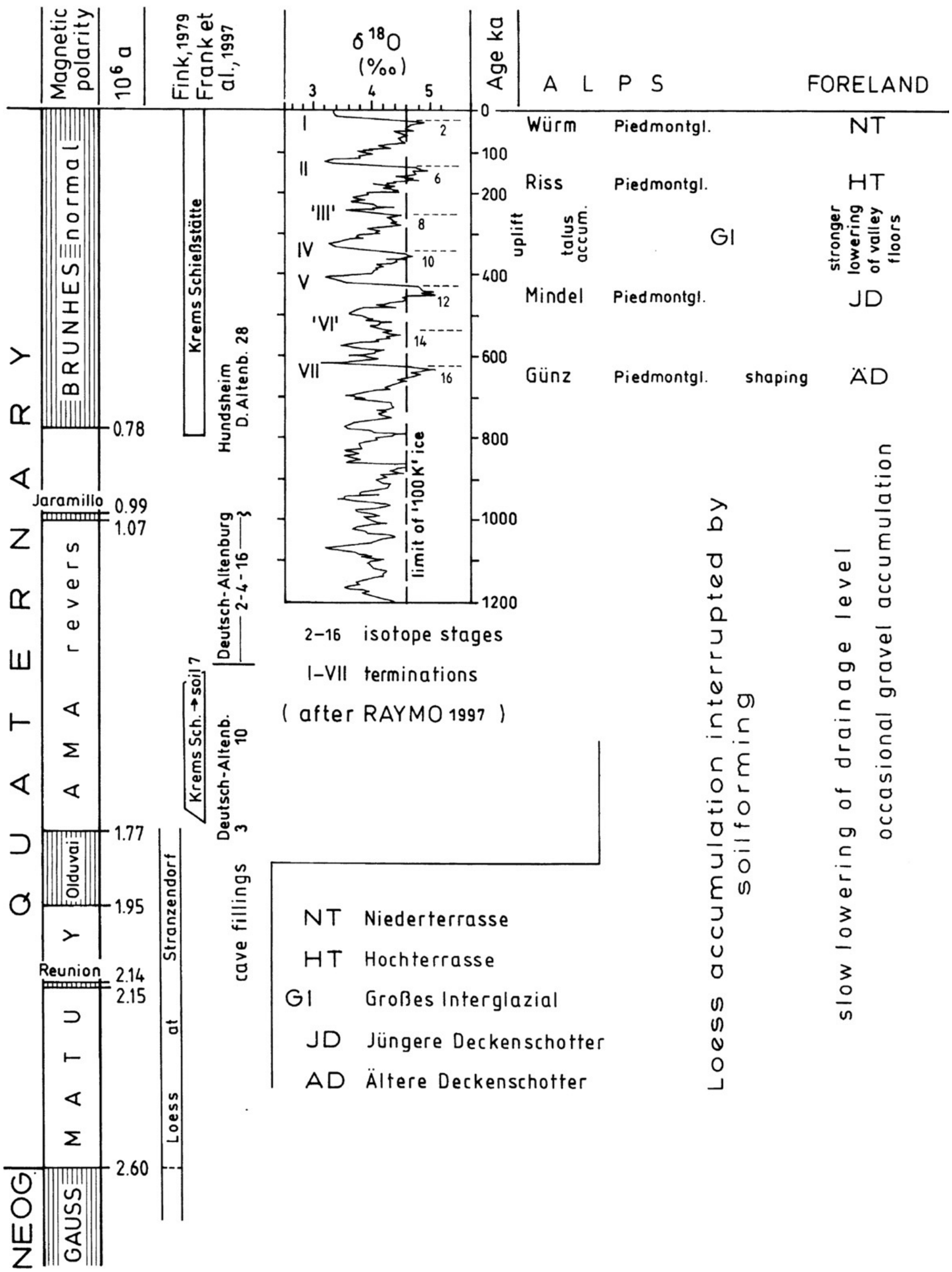

FIGURE 2.1 Geological time scale of the Quaternary events in Austria.

Palaeomagnetic investigations (Fink, 1979) suggest that the tills of these glaciations and their connected outwash terraces all fall into the Brunhes Chron. The marine oxygen isotope record suggests that four major glaciations affected the northern hemisphere during the Brunhes Chron (Shackleton, 1987; Raymo, 1997), generally correlated with MIS

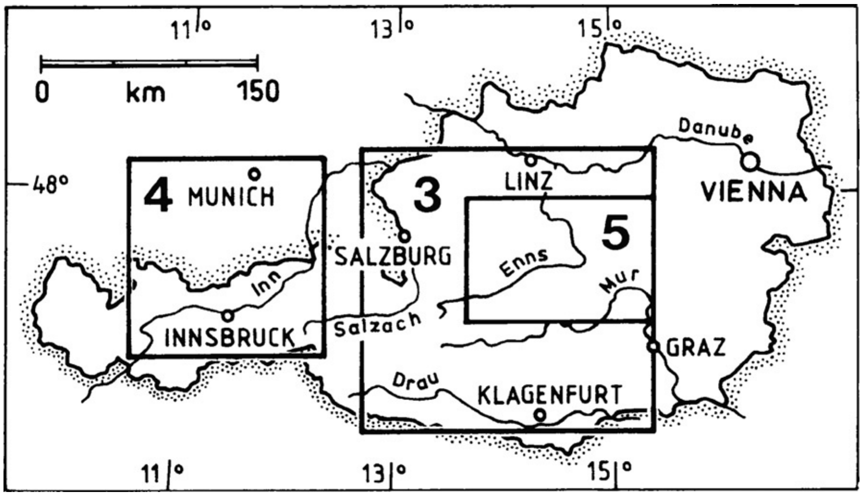

FIGURE 2.2 Location map showing the main drainage systems of Austria. Squares mark figures 2.3, 2.4, 2.5.

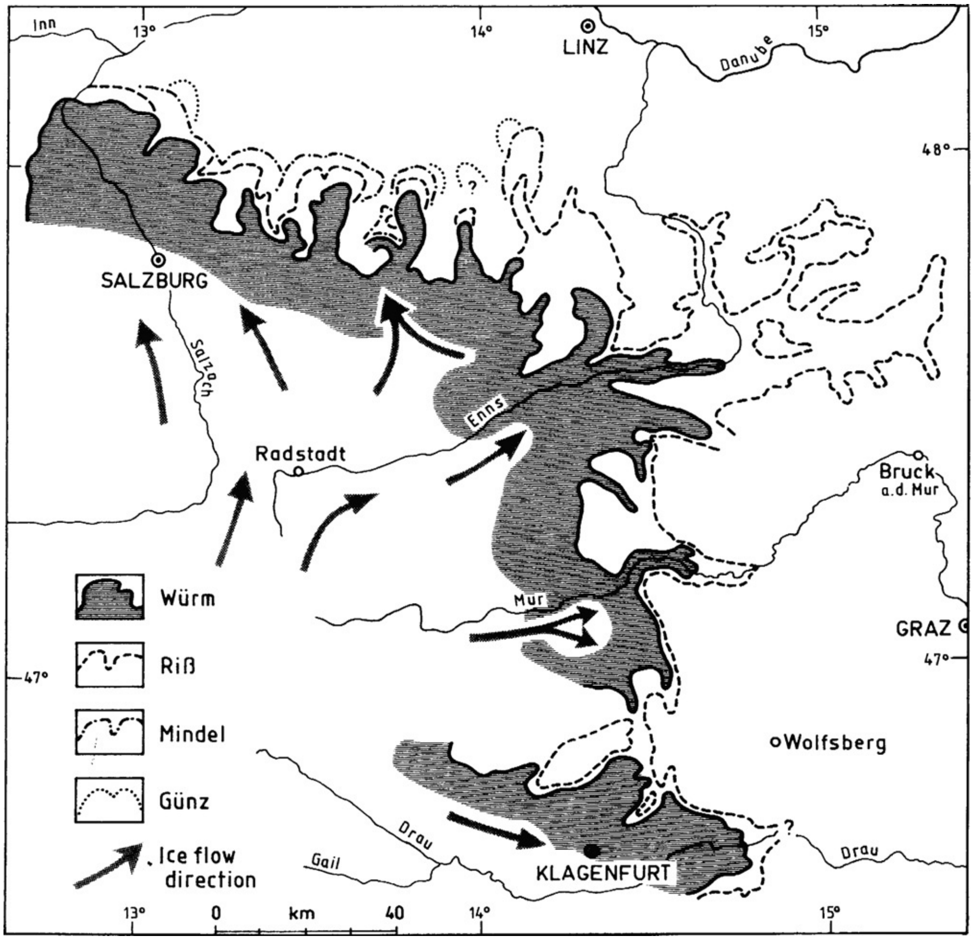

FIGURE 2.3 The extent of the glaciers of the four glaciations known in Austria. Only glaciers of the main valleys are indicated. Ice flow direction is rather similar during each glaciation.

2, 6, 12 and 16 (Raymo, 1997). According to radiometric dating, stratigraphical position as well as weathering, the two youngest Alpine glaciations are correlated with MIS 2 and 6. However, the much stronger weathering, cementation and periglacial modification of all the Mindelian deposits, in comparison to those of the Rissian Stage, indicate that a longer time span must have separated these glaciations. This *Großes Interglazial* (great interglacial) was first postulated by Penck and Brückner (1909/11). Thus, the Mindelian probably correlates with MIS 12 and the Günzian with MIS 16 (Fig. 2.1).

Remnants of another major cold period with gravel deposition in the Alpine foreland that fall between depositions of the Günzian and Mindelian stages have been found more recently (Kohl, 1976). They are separated from both by weathering. This cold stage probably correlates with MIS 14. Extents of the glaciers of these two oldest glaciations were controlled by partly different valley drains leading to locally singular and longer toungues (Fig. 2.3).

2.4. TECTONIC ACTIVITY

The incision of the circum-alpine drainage system as far as the level of the *Ältere Deckenschotter* and beyond seems to have been of an intensity similar to that observed in the Danube and its tributaries (Graul, 1937; Fuchs, 1972; Fischer, 1977). This suggests that no differential uplift has occurred and that there are no traces of other tectonic activity, such as faulting within the northern foreland of the Alps. However, age control of the remnants of terrace systems older than the *Ältere Deckenschotter* is not yet possible.

On the basis of the loess sequence near Krems (Fink et al., 1976), the amount of erosion along the Danube system was about 50 m during the Quaternary period (Fig. 2.1). This supposedly undisturbed development along the northern edge of the Alps ends where the Danube enters the Vienna Basin. Clear evidence of tectonic activity that continues well into the Holocene is recognisable here (Plachy, 1981). The glaciofluvial terraces dating from the last two glaciations (*Hochterrasse, Niederterrasse*, as well as Holocene deposits) show clear evidence of tectonic displacement; indeed parts of the *Hochterrasse* are displaced by as much as ca. 10 m along major faults (Kröll et al., 1993; Decker et al., 2005).

Traces of recent tectonic activities in Quaternary sediments within the Eastern Alps, as would be expected from Miocene and Pliocene development (Peresson and Decker, 1996), have yet not been proved. Only along the rivers Enns and Steyr, where glaciers have not obscured evidence, a displacement of the *Jüngere Deckenschotter* at the northern edge of the Alpine nappe system indicates a tectonic activity before the penultimate glaciation (van Husen, 2000). This may have been associated with and caused by an uplift of the whole mountain chain (Fig. 2.1).

2.5. DEVELOPMENT OF GLACIERS

The Alpine drainage system was filled by dendritic glaciers during all four major glaciations. The topography of the longitudinal valleys (e.g. the Inn, Salzach, Enns, Mur and Drau; Fig. 2.2) controlled glacier build-up in their tributaries and surrounding areas. On the one hand, this explains the outstanding differences in glacier extent between some of the catchment areas. On the other, it explains the strongly accelerated ice build-up towards the glacial maximum. Two examples are given here.

2.5.1. Inn Valley

The Inn valley is the most extensive drainage system in the Eastern Alps. Its source areas are located south of the valley and in the Engadine, areas almost exclusively underlain by crystalline rocks. North of the Inn valley, the Northern Calcareous Alps have no major valley tributary of the Inn, which follows the border between these two tectonic units east of Landeck (Fig. 2.4).

The valleys in the Northern Calcareous Alps are orientated to the north (Fig. 2.4), forming the heads of the rivers Isar and Loisach and other small rivers. The Inn valley is separated from them by mountain chains rising up to 2500 m a.s.l. or more. Three gaps in the mountain chain are found at Fernpass, Seefeld and Achensee, with watersheds 400–600 m above the Inn valley. During the major glaciations, these cols were crossed by ice (Fig. 2.4). A similar behaviour can also be reconstructed by erratic material in basal tills in the valleys of Salzach, Enns, Mur and Drau (see map).

Till composition reflects the petrographic composition of the catchment areas. In the basal tills of the Inn valley, crystalline components are dominant; only on the northern slope is there a considerable admixture of limestones, derived from the local bedrock. However, up to 35% of crystalline erratics (Fig. 2.4) are found in the glacial and glacigenic material of the Würmian Isar and Loisach glaciers (Dreesbach, 1983). Most of the crystalline material is fresh and unweathered except for a small amount affected by multiple reworking. Such a large amount of unweathered crystalline rocks has to be attributed to extensive ice transport from the Inn valley to the Isar and Loisach systems, across the Fernpass and the Seefeld passes (Penck and Brückner, 1909/11). This transfluence reflects the gradient of the ice surface. The ice filling the Inn valley reached a much higher elevation than in the valleys in the north. The extraordinarily high crystalline boulder content in the sediments in the advance phase of the ice streams suggests a strong and rapid ice build-up in the Inn valley between

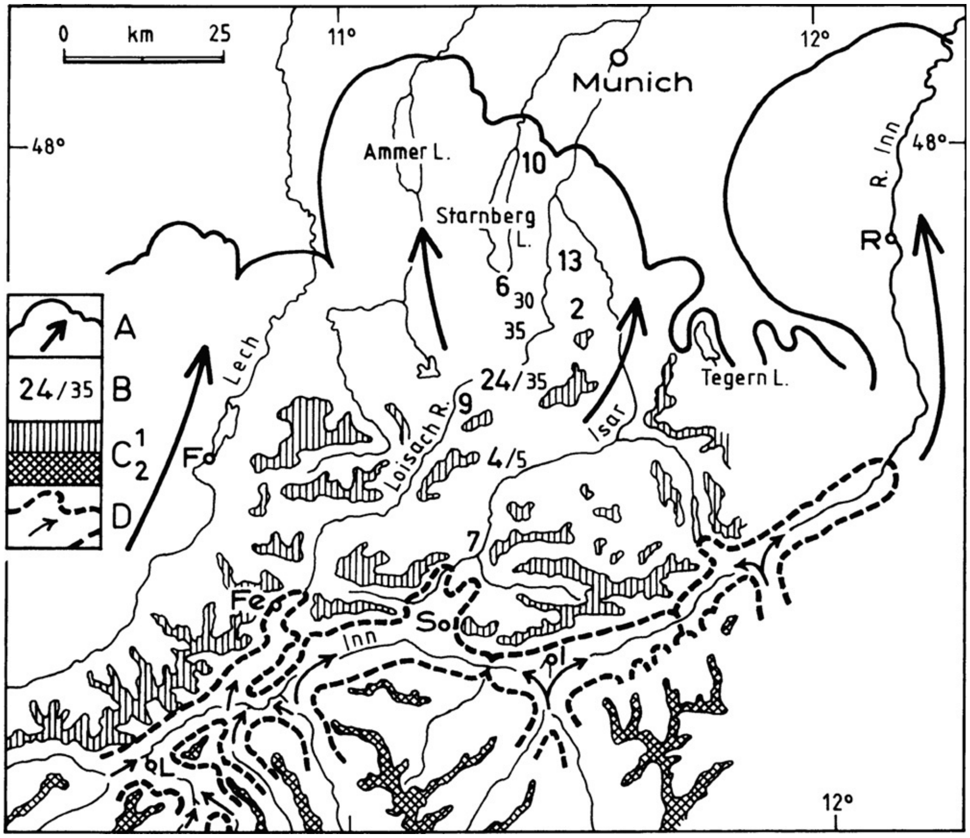

FIGURE 2.4 Ice streams of the Inn, Lech, Isar and Loisach valleys. (A) Glacier extent during the Würm MIS 2. (B) Percentage of crystalline boulders. The larger numbers in basal till. (C) Nunataks formed of sedimentary rocks (C1) and crystalline rocks (C2). (D) Probable ice extent in the Inn valley around the beginning of the final ice build-up during MIS 2. F: Füssen, Fe: Fernpass, I: Innsbruck, L: Landeck, R: Rosenheim, S: Seefeld.

Landeck and Innsbruck. This would allow strong transfluence from the Inn valley even at an early phase of the glaciation, suppressing the influence of the local glaciers and affecting both till and outwash gravel composition. Such an early overflow may result from the internal ice flow conditions within the Inn valley system. The large tributaries from the south, with their extensive source areas at high altitudes, carried huge glaciers during the final approach of the last glaciation, gradually filling the whole Inn valley. Hence, five glaciers finally merged around Landeck. At the same time, the ice flowing out of the Sill and Ziller valleys reached the main valley and thus the glaciers may have blocked each other creating a conspicuous rise in the ice surface. Therefore, crossing the watersheds, the ice of the Inn valley system reached the drainage systems of the Isar and Loisach (with smaller catchment areas at lower altitudes) early enough before these valleys were filled with local ice strong enough to hinder this process.

However, ice congestion in the narrow Inn valley also resulted in a high ice table leading to a rapid expansion of the feeding area, which again favoured rapid ice build-up. Thus, the high gradient to the north persisted throughout the entire glaciation. The volume of ice discharge north across the watersheds (crystalline material included) can be judged from the size of the Isar and Loisach piedmont glaciers as compared with their neighbours, which were fed only by their own source areas within the Northern Calcareous Alps (Fig. 2.4).

2.5.2. Enns Valley

The Enns valley glacier (Fig. 2.2) was connected to the Salzach glacier in the west and the Traun glacier in the north (Fig. 2.5). The Enns valley follows the border between the crystalline zone in the south and the Northern Calcareous Alps in the north. The topography differs slightly from that of the Inn valley. In the western part, the tributaries from the south originate in extensive areas of high elevation, and the wall-like Northern Calcareous Alps drain mainly towards the north. Further to the east, however, the elevation of the crystalline zone declines abruptly over a short distance, and the Northern Calcareous Alps changes from continuous chains and plateaux to more isolated mountains surrounded by much lower terrain.

During the Würmian, the Enns valley was occupied by a glacier extending down to the area of more isolated high

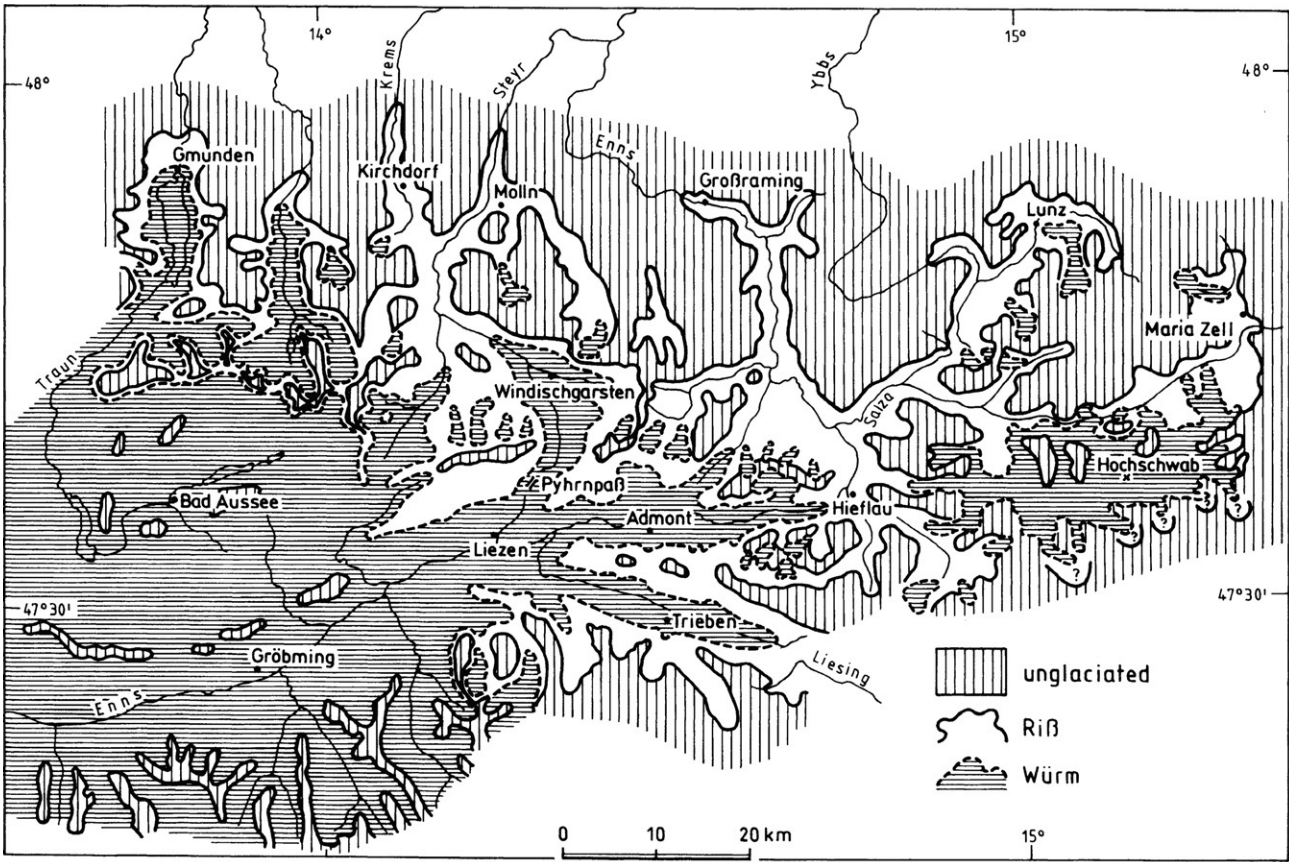

FIGURE 2.5 Glacier extent during the Würmian (MIS 2) and Rissian (MIS 6) in the Enns valley and environs and in the Steyr and Ybbs valleys.

mountains. The latter supported local glaciers (Fig. 2.5). Both the local glaciers and the valley glacier were in contact but had little influence on one another with regard to the ice discharge. To the south, a 20 km long tongue of the valley glacier entered the otherwise unglaciated valley around Trieben. To the north, ice crossed the Pyhrn Pass, entered the drainage system of the Steyr river and filled the small basin at Windischgarsten. The content of crystalline boulders in the tills is around 2–3% in this area. East of the Enns glacier only more or less small local glaciers developed. Their extents depended strongly on snowdrift by prevailing westerly winds.

During the penultimate glaciation (Rissian), the equilibrium line of the glacier system was about 100 m lower than during the Würmian (Penck and Brückner, 1909/11). This led to more extended piedmont glaciers on the northern rim of the Alps and longer valley glaciers within the mountains. The difference in length was in most cases about 5–6 km (Fig. 2.3). The situation was very different in the Enns and Salza valleys and the Steyr-Krems drainage system (Fig. 2.5). In the former, the end moraines at Grossraming (van Husen, 1999) show a glacier extending ca. 40 km further than during the Würmian. The tills throughout the valley and in the immediate vicinity north of Hieflau contain abundant crystalline boulders, indicating that the Enns valley was filled by a much larger Enns glacier with little influence from the local glaciers.

With respect to the lower position of the equilibrium line during the Riss glaciation, both the local glaciers and the valley glacier grew much larger. Thus, the local glaciers were powerful enough to block the valley glacier in the very narrow portion of the valley west of Hieflau. This impediment to ice flow caused a higher ice surface in the Enns valley west of Admont. This area, that formed part of the ablation area during the Würmian, consequently became an accumulation area in the Rissian Stage. This considerable addition to the accumulation area, together with the ice supply from the local glaciers, affected the extent of the glacier, as explained above.

The glaciers on the north slope of Hochschwab grew in the same way (Fritsch, 1993; Kolmer, 1993), filling the Salza valley, which became part of the accumulation area. As a result, an extended ice stream formed here. This process may have been accelerated by the additional cooling effect of the glaciated Enns valley (Fig. 2.5).

The same feedback mechanism occurred in the Windischgarsten basin. The stronger transfluence of ice from the Enns valley in the south caused a larger glacier tongue to develop and to extend further and further into the basin. In the same way, the larger local glaciers contributed to the overall ice volume. Thus, the whole basin became part of the accumulation area. This greatly enlarged feeding area enabled the glacier to advance down valley over 30 km beyond its Würmian limits. It was even powerful enough to cross the watershed into the Krems river system (at Kirchdorf on the Krems), generating a much greater ice stream in that valley (Kohl, 1976).

2.6. OVERDEEPENED VALLEYS

Since the beginning of research on ancient glaciers and their palaeogeographical distribution, overdeepened tongue basins have been known (Fig. 2.6). They seem to form predominantly in the ablation area, where the higher ice velocity increases the basal debris load and where basal meltwater drainage under hydrostatic pressure occurs (van Husen, 1979). They were shaped successively during each glaciation because the tongue areas always developed more or less in the same positions (cf. Fig. 2.3). Hydrogeological investigations in the longitudinal valleys provide new data on the shape and depth of these basins. Geophysical investigations, together with boreholes, also provide good evidence on the sediment filling of the basins and the position of the underlying bedrock.

Thus, on the one hand, the underlying bedrock in the tongue basins of the Salzach glacier was cored repeatedly between 160 and 340 m, on the other, the base in the Gail and Drau valleys was found at 200–240 m (Kahler, 1958). These investigations did show that the overdeepening may be limited to about 400 m in this part of the Austrian Alps. Similar depths are also found in some lakes (e.g. Traunsee), which have a depth of nearly 200 m and a thick sediment fill at the bottom. A depth of ca. 200 m was also determined for the Steyrtal, which was affected by the most extensive glaciations (Fig. 2.6, M) (Enichlmaier, personal communication).

Stronger overdeepening is reported from the Inn valley. East of Innsbruck, for example, the pre-Quaternary basement lies at ca. 180 m a.s.l., about 400 m below the valley floor (Aric and Steinhauser, 1977). Further to the east, seismic investigations revealed that the bedrock lies at about 500 m below sea level (1000 m below the valley floor). The latter was proved by a drill hole 900 m deep that passed through unconsolidated gravel, silt and clay without reaching the bedrock (Weber et al., 1990). This excessive amount of erosion and overdeepening may result from the stronger linear ice and meltwater discharge along valleys with

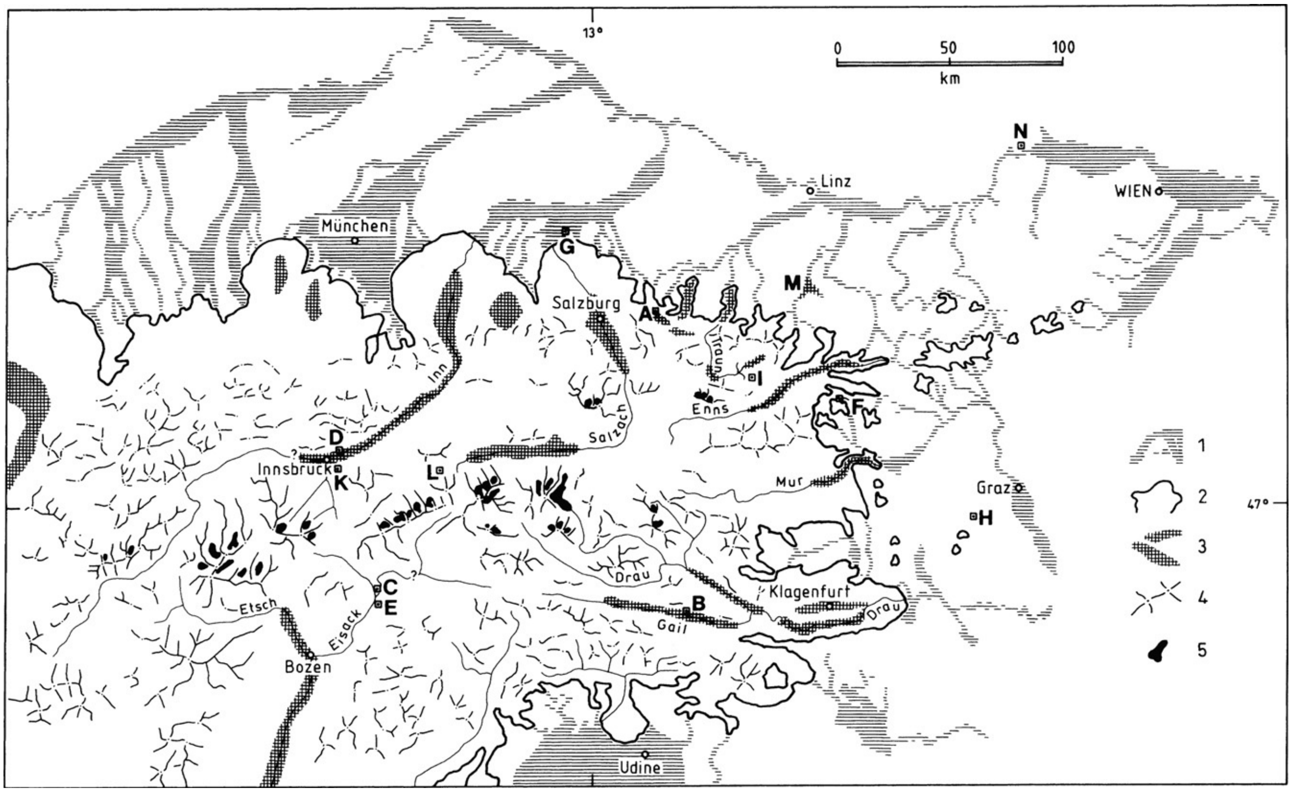

FIGURE 2.6 Sketch map of the Eastern Alps during Würmian (MIS 2); 1 Terrace 'Niederterrasse'; 2 Maximum extent of glaciers; 3 Overdeepened parts of the valleys; 4 Nunataks; 5 Glacier extent of the Holocene. Localities mentioned in the text: A: Mondsee; B: Nieselach; C: Schabs; D: Baumkirchen; E: Albeins; F: Hohentauern; G: Duttendorf; H: Neurath; I: Mitterndorf; K: Lans; L: Gerlos; M: overdeepened area at Molln.

extensive catchment areas. It is in close accordance with the erosion of about 600–1000 m, recently reported from the longitudinal valleys of Rhine and Rhône in Switzerland (Pfiffner et al., 1997).

The filling of the overdeepened valleys depends strongly on the relation of the major rivers to their tributaries, in terms of water and debris discharge, as well as on the size of the basin. Large basins, with a strong main river and small tributaries, were often filled with a thick sequence of fine-grained bottom-set beds that interfinger with coarse delta deposits (e.g. Salzachgletscher: Brandecker, 1974). Strong input of coarse gravel and sand from major tributaries creates a more inhomogeneous valley fill with alternating layers of gravel, sand and silt all over the area.

2.7. LAST INTERGLACIAL-GLACIAL CYCLE

According to oxygen isotope records, all climatic cycles within the Brunhes Chron and in the Late Matuyama Chron show a similar pattern. This is particularly true of the four major cycles before the Terminations I, II, V and VII, following the major glaciations at MIS 2, 6, 12 and 16 (Raymo, 1997) each of which is characterised by a step by step cooling, interrupted by short phases of climatic amelioration, leading eventually to the very cold short true glaciation period (Fig. 2.1). The last, Eemian/Würmian, climatic cycle is relatively well investigated. Therefore, it may serve as a model for the reconstruction of the other cycles in terms of climatically induced sedimentation and facies diversification in the Eastern Alps (van Husen, 1989).

On the northern edge of the Eastern Alps (Fig. 2.6, A), the Mondsee sedimentary sequence has given good, continuous evidence on the climatic development from Termination II, through the Eemian, and well into the first half of the Würmian Stage. The fine-grained sediments, north of the shoreline of the present lake Mondsee, were first investigated by Klaus (1987) and believed to represent a complete sequence covering the time between the Rissian and Würmian glaciations. Recent investigations, based on three long cores, have revealed a delta structure in an ancient lake with a water surface around 50 m above the present lake level. Bottom set, foreset and thin topset beds of a classical Gilbert-delta structure were covered by the till of the last glaciation (Krenmayr, 2000).

The results of palynological investigations (Drescher-Schneider and Papesch, 1998; Drescher-Schneider, 2000) suggest that the Eemian in this area was a warm period with temperatures averaging 2–3 °C above the current Holocene values. The valleys and foreland of the Alps were densely forested at this time with well-developed mixed oak forest with a high content of *Abies* (fir). This phase ended with an abrupt climatic deterioration that affected the forest on the northern edge of the Alps and brought coarser sediments to the delta.

During the Early Würmian, forests recovered twice, showing some elements of mixed oak forest, with a cold stadial intervening between these periods. During the cold stadial, the treeline descended close to the lake level. The cold period at the beginning of the Middle Würmian saw the treeline more or less at the level of the Alpine foreland. This was followed by a slight warming, allowing a forest dominated by *Larix* (larch) and *Picea* (spruce) to grow around the lake. This series of events corresponds closely to the Samerberg sequence from east of the Inn valley (Grüger, 1979), as well as with that at La Grande Pile in the Vosges (Woillard and Mook, 1982).

2.8. CHRONOLOGY

The final climatic deterioration and glacier advance phase is represented at the Würmian Albeins and Baumkirchen (Fig. 2.6, D and E) sites where till overlies gravel and lake deposits, respectively. Radiometric dating indicates that the glaciers of the tributary valleys reached the main longitudinal valleys at about 25–24,000 years BP. The rate at which further ice build-up occurred is unknown because of the total lack of chronological evidence. During this final climatic decay (Fig. 2.8), the valleys were filled with coarse gravel, *Vorstoßschotte* at many places to high elevations, as a result of progressive overloading of the main river with debris (van Husen, 1989). The *Vorstoßschotter* extend laterally into the terraces in the foreland, which accumulated at the same time as the thick gravel bodies along the rivers (Fig. 2.7). These were developed also in non-glaciated areas by periglacial activity (congelifraction). In only two places, Duttendorf and Neurath (Fig. 2.6, G and H), in the Eastern Alps, can radiocarbon dates constrain the climax of the climatic deterioration and the maximum extent of glaciers and periglacial activity, as well as strong congelifraction and periglacial downwash around 20,000 years BP (van Husen, 1989).

Around the Eastern Alps, detailed mapping of terminal moraines and outwash terraces (e.g. the Traun, Enns, Mur, Drau) has shown that most glaciers behaved in a similar way. First of all, the greatest extent of the glacier tongues is marked by small morainic ridges connected to outwash fans. After this, the glacier fronts retreated some hundreds of metres and formed distinctive, high and wide end moraines, also connected to outwash, which grades morphologically into the downstream fluvial terraces. These outwash fans, and their transition into downstream terraces, have a lower gradient than the earlier ones. Within 1–2 km, they are on the same level, merging to form the *Niederterrasse* (Lower Terrace), which continues downstream and can also be traced along the Danube to Vienna (van Husen, 1987). The terraces also correspond to those of the unglaciated tributary valleys.

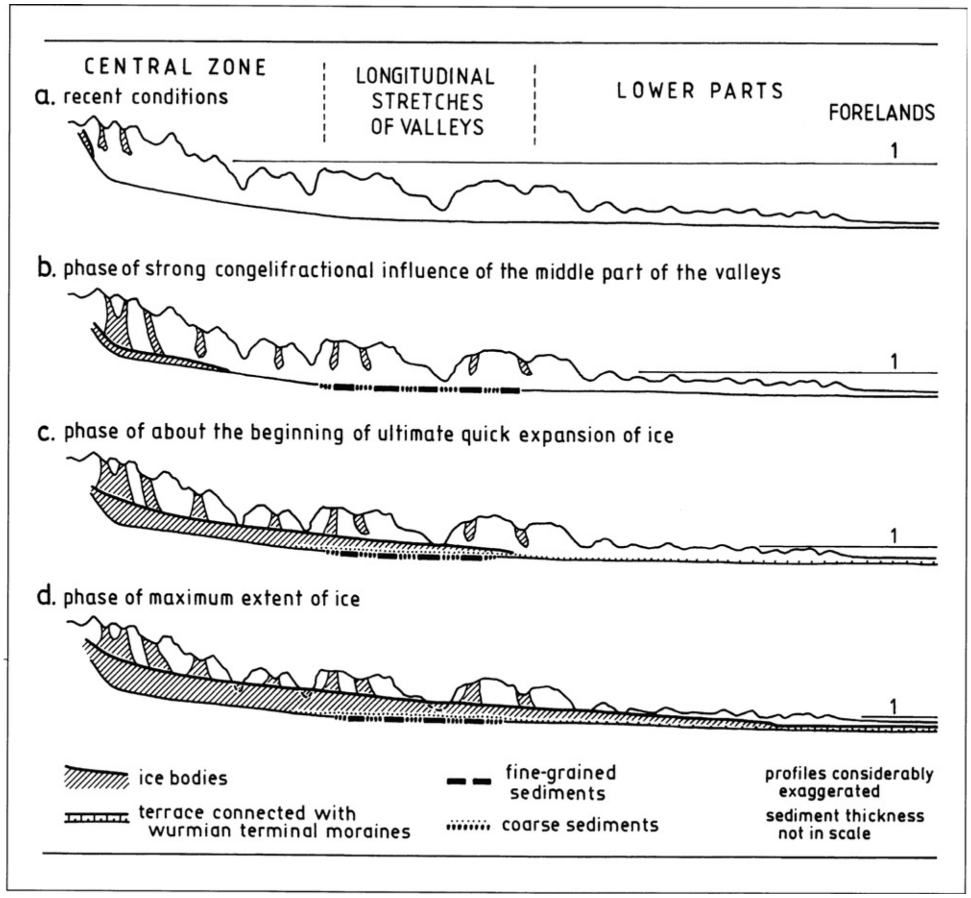

FIGURE 2.7 Sketch profiles showing the consequences of gradual climatic deterioration indicated by the estimated lower limit (l) of strong congelifraction. It shows that only a little lowering of this limit and the equilibrium line (C to D) is necessary to cause a very rapid and substantial expansion of the valley glaciers, such as during the Würmian maximum (see Fig. 2.8).

Evidence of weathering has been found neither between the sediments of these different Würmian terminal moraines nor within the outwash gravel sequence. This suggests that very little time elapsed between these two events, the *Maximalstand* and the *Hochstand* (van Husen, 1977, 2000). It is not known how long the glaciers were in place to form the 20–40 m high moraines of the *Hochstand*, as no datable material has yet been found. An identical differenciation was recently observed at the Tagliamento where organic material allowing to determine time slots of 26.5–23.0 and 24–21 cal. ka BP for these events (Monegato et al., 2007).

The first retreat from these terminal moraines was in the order of some hundreds of metres to several kilometres, depending on the size of the glacier. This generally led to a drainage concentration to only one or two outlets, and an initial minor incision into the outwash fan and terrace. This stage is characterised by small morainic ridges and kettle holes, indicating continuing permafrost conditions. Large blocks of ice were preserved below the sediments during the whole time span from the glacial maximum to its early retreat phase (van Husen, 1977). The kettle holes formed after the downmelting of the glacier tongue, when a single deeply incised outlet drained the overdeepened glacier basin.

The duration of the maximum extent of glaciation and climatic deterioration during the LGM (Last Glacial Maximum) can only be tentatively estimated. After the ice advance around 21 ka BP, the glacier front may have remained in these three maximal positions for about 4000 years, on the basis of data from the subsequent deglaciation phases.

2.9. PHASE OF ICE DECAY

Following the LGM, large-scale retreat and downwasting of the glacier tongues began in the Alpine foreland, as well as in the valleys. In all the great valleys of the Eastern Alps (Inn, Salzach, Drau, Mur, Enns and Traun), no sequences of end moraines or equivalent evidence of former ice margins have been found from this phase. Only kames and ice-marginal terraces that formed in temporary lakes have been identified. These ice-contact sediments,

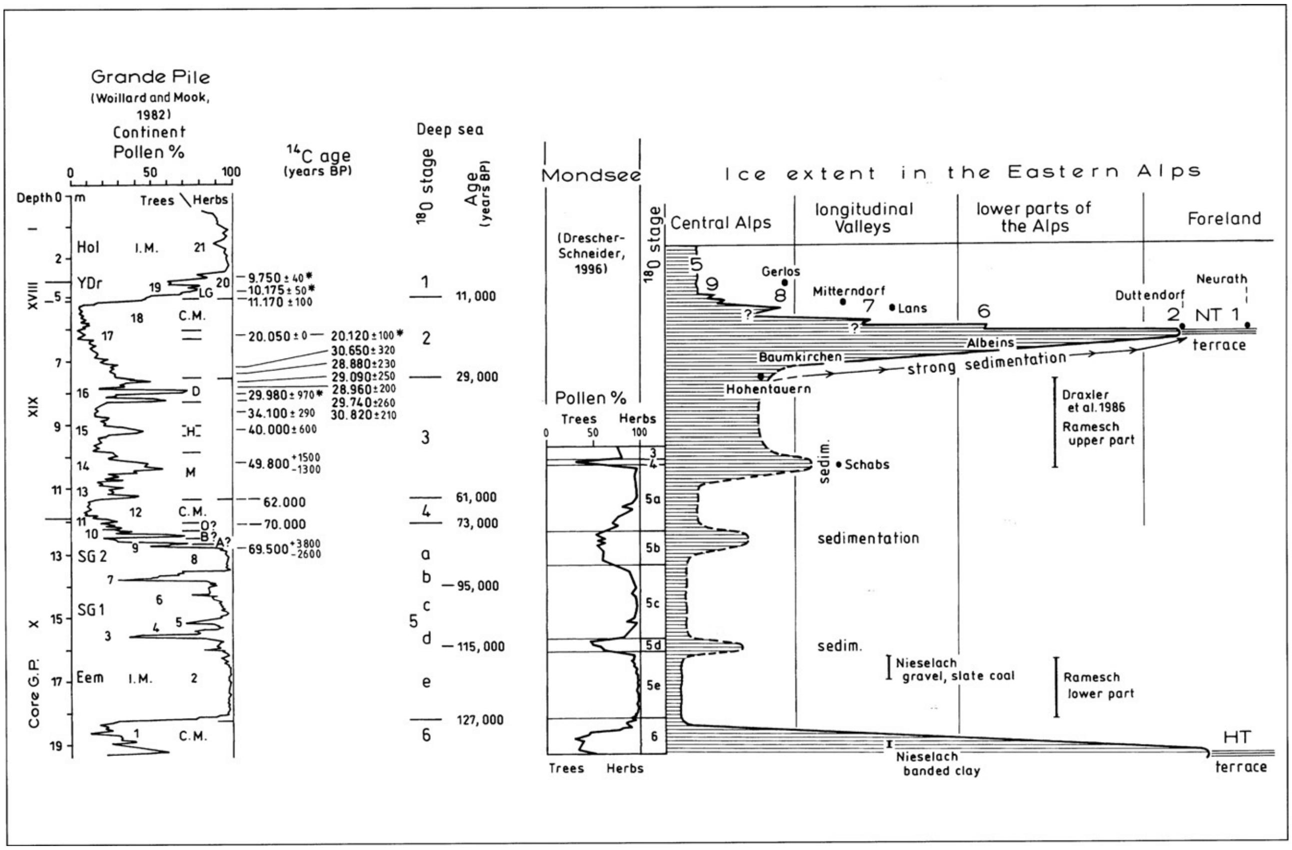

FIGURE 2.8 Temporal development of the ice extent in the Eastern Alps over the past 140,000 years (after van Husen, 2000).

developed especially around the overdeepened parts of the valleys (Fig. 2.8), indicate continuous downmelting, without glacier stillstands or readvances. With respect to the distribution and internal structure of the sediments which formed in temporary lakes, downmelting was rapid. Some hundreds to one thousand years may have been all that was necessary for the loss of about 50% of the glacier lengths in the Eastern Alps. Ice lakes first formed at the glacier margins and then extended over the entire area of the overdeepened basins. This probably resulted in glacier calving, which would have enhanced the rate of ice recession.

Knowledge on the further recessional phases is mainly based on investigations in two valley systems. One is the Traun valley, where the complete sequence of retreat and readvance phases from the LGM to the beginning of the Holocene was mapped in this relatively short valley and investigated by sediment analysis, palynology and radiocarbon dating (Draxler, 1977; van Husen, 1977). A second is the Inn valley where all the type localities of these events are situated (Mayr and Heuberger, 1968), and where intensive investigations have recently been undertaken by Bortenschlager and Patzelt. The close correspondence of the sequences in both areas, in terms of sediment and vegetation development and radicarbon dating, allows the use of classical terms for easier understanding. The following paragraphs describe key sites that provide some evidence about the deglaciation of the Eastern Alps.

2.10. THE BÜHL PHASE

The first sign of a halt in the downmelting of the glaciers is marked around the intramontane basin of Bad Ischl. Here extensive kame deposits, partly covered by a thin layer of till, are connected to small morainic ridges, and this assemblage suggests a stillstand of the glacier margin with minor oscillations (van Husen, 1977). This phase is comparable to the 'Bühl Stage' of Penck and Brückner (1909/11), as shown by the more detailed investigation of the type locality by Mayr and Heuberger (1968). Recent geological re-investigations of the type locality revealed no remarkable readvances there. Predominantly, the forming of extended kame terraces connected with stagnant and downmelting icebodies probably occurred during the period of the Greenland Stadial 2c 21.2–19.5 ka BP (Reitner, 2007). The lithology of pebbles and boulders in the till suggests that a dendritic ice stream in the main valley was still connected to glaciers in all of the tributary valleys at this time. Large kame terraces also exist in other valleys, such as the Drau valley, but they have not been mapped and studied in detail. Nevertheless, their general distribution suggests that they were associated with ice streams comparable to those typical of the Bühl Phase (Fig. 2.8).

2.11. THE STEINACH PHASE

The phase of deglaciation that followed the Bühl was characterised by a minor readvance of the by then much smaller glaciers, again linked with kame terraces and inactive ice masses. Thus, the glacier tongue in the Traun valley near Bad Goisern had advanced over lacustrine and fluvial sediments deposited high above the valley bottom, apparently formed when drainage in the valley to the north was still blocked by stagnant ice masses (van Husen, 1977). A similar situation was described for the Steinach Phase (a term introduced by Senarclens-Grancy, 1958) from the Sill valley, south of Innsbruck (Mayr and Heuberger, 1968). Here, a thick sequence of gravels was also deposited in contact with stagnant ice masses and was covered by the till of an ice readvance. Thus, after the Bühl Phase, the main valleys had become free of active ice and the glaciers had retreated into the tributary valleys. The time interval between the two phases could not have been very long, as large inactive ice masses lingered in the valleys, despite climatic amelioration and large meltwater lakes (Fig. 2.8 and 2.9).

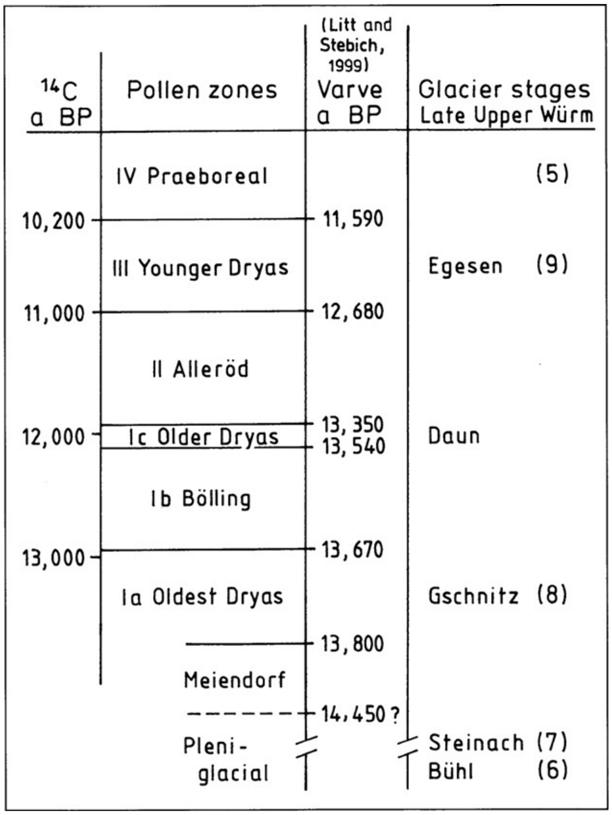

FIGURE 2.9 Temporal position of the late Upper Würmian phases (Termination 1). For explanation see Fig. 2.8.

2.12. THE GSCHNITZ PHASE

The next phase in the deglaciation sequence is marked by well-developed blocky end moraines. In the Traun valley, these are present around Bad Goisern and in all of the other source areas of the Würmian ice stream. The morainic ridges are connected to outwash gravels almost everywhere. In the Traun valley north of Bad Goisern and around Bad Aussee, these deposits form terraces extending about 10 km downstream. This indicates that the valleys were free of dead-ice, permitting free drainage along the valley bottoms (van Husen, 1977). This suggests a comparatively long period of time, probably climatic amelioration between the Gschnitz and Steinach Phases and considerable ice recession (Fig. 2.8). The Gschnitz moraines are relatively unmodified by slope processes suggesting that little or no solifluidal shaping has occurred since their deposition. The moraines of the Steinach Phase, in contrast, were clearly smoothed by solifluction during the following (Gschnitz) glacial event. This Gschnitz Phase is also well developed at Trins, south of Innsbruck (Mayr and Heuberger, 1968). Moreover, similar distinctive moraines can be recognised in many of the large tributary valleys draining from higher parts of the central Eastern Alps, as well as in the high cirques to the north and south. This implies that the glacial event was regionally extensive, reflecting a uniform lowering of the equilibrium line (ELA) to a position about 600 m lower than that of the Little Ice Age (Gross et al., 1978).

2.13. CHRONOLOGY

These glacial phases have been indirectly dated by palynological studies of bogs in the Traun valley (Draxler, 1977, 1987). During the early phase, after the main LGM deglaciation, some depressions were slowly filled with varved clay. The pollen record of this time is dominated by *Artemisia*, *Helianthemum*, *Ephedra*, *Hippophaë* and *Juniperus*, in addition to *Pinus*; the *Juniperus* becomes important towards the end of the sequence. This vegetational assemblage, especially the high content of *Artemisia*, is typical of the pioneering phase under dry, cold conditions (Draxler, 1987). The same feature of this early phase has also been described from the western part of the Eastern Alps (Bortenschlager, 1984). The phase terminates with the rapid increase of *Pinus* pollen to values of 70–80%. This interval is well dated in the Traun area to around 12.3 ^{14}C ka BP. The following dates are reported in van Husen (1977):

Moos Alm: 730 m a.s.l., 12.52 ± 0.18 ^{14}C ka BP. VRI-431;

Ödensee: 770 m a.s.l., 12.22 ± 0.18 ^{14}C ka BP. VRI-433;

Plakner: 550 m a.s.l., 12.41 ± 0.19 ^{14}C ka BP. VRI 430; Ramsau: 515 m a.s.l., 11.97 ± 0.2 ^{14}C ka BP. VRI-432; Rödschitz: 790 m a.s.l., 12.44 ± 0.42 ^{14}C ka BP. VRI-485.

Dates for the equivalent interval from the Tyrol, reported by Bortenschlager (1984), include:

Lanser See: 840 m a.s.l., 13.23 ± 0.19 ^{14}C ka BP. HV 5269, and
Gerlos: 1590 m a.s.l., 12.155 ± 0.21 ^{14}C ka BP. HV 5284.

The difference of some 100 years between sites may be due to the contrasting rates of soil-forming processes on limestone and crystalline bedrock, as well as to differences in plant immigration. However, the dates suggest that this event occurred during or at the end of the Bølling Ib Chronozone (Fig. 2.9). Bogs documenting this event in the Traun valley lie both outside and inside end moraines of the Gschnitz Phase. Thus, this glacial advance occurred no later than the Oldest Dryas.

This event can be dated more precisely in the pollen profile from Rödschitz in front of the Gschnitz moraines, in which the cooling event is marked by a strong increase in *Artemisia* at 6.40 m depth. Radiocarbon dates from gyttja at 5.40 m depth (12.42 ± 0.44 ^{14}C ka BP. VRI485) and organic detritus (pieces of shrub and herbs) at 7.20–7.00 m depth yield an age of 15.4 ± 0.47 ^{14}C ka BP. VRI-484 suggest that the Gschnitz cold phase had occurred around 14 ^{14}C ka BP, assuming an approximately constant rate of sedimentation in the lake. A similar estimate was made by Patzelt (1975).

Recently, it was attempted to date this event by surface exposure dating (^{10}Be and ^{26}Al) at the type locality (Ivy Ochs et al., 1997, 2000) indicating a final forming of the terminal moraine before 16 ka (Ivy Ochs et al., 2006). This probably would push it into a temporal position before the Meiendorf Stadial at the end of the Pleniglacial (Fig. 2.9).

The periglacial modification of end moraines of the earlier Steinach advance, and the lack of such reshaping on the Gschnitz moraines, agrees well with an Oldest Dryas age for the Gschnitz event, immediately preceding the climatic improvement at Termination l. After the Bølling Interstadial, there is no evidence for permafrost conditions on the floors of the main valleys in the Eastern Alps.

Based on evidence from the Rödschitz site (basal date of ca. 15.4 ^{14}C ka BP), the Steinach event may have occurred at around 16 ^{14}C ka BP (Fig. 2.9). Thus, the earlier Bühl Phase possibly culminated shortly before this date. During the warmer conditions of the Bølling Chron, the valley bottoms became ice free. Only high parts of the limestone plateaux of the Northern Calcareous Alps and valleys at higher elevations in the Central Alps remained covered with ice.

2.14. THE DAUN PHASE

During the subsequent short, cold Older Dryas Chron, enhanced ice accumulation stimulated small glaciers on the higher limestone plateaux, like the Dachstein Plateau. These small ice masses formed blocky end moraines, including boulders the size of houses. However, in general, this and the following (Egesen) event are not well marked on the comparatively low terrain of the limestone plateaux. By contrast, in the higher parts of the Eastern Alps, moraines of Older and Younger Dryas age are well developed. The older event shows the ELA at a position more than 300 m lower than that of the AD 1850 (Little Ice Age) snowline (Gross et al., 1978).

2.15. THE EGESEN PHASE

This event is marked by well-developed end moraines, which, according to palynological records and radiocarbon (Patzelt and Bortenschlager, 1978) and exposure dating (Ivy Ochs et al., 1996), are believed to have formed during the Younger Dryas Chron of NW Europe. The ELA was lowered by about 300 m (Gross et al., 1978) at this time, but arising from precipitation differences across the mountains, this value varied between 280 m in the drier continental part and 400 m in the more oceanic northern ranges (Kerschner, 1980). Generally, the Younger Dryas was characterised by drier, more continental conditions (around 70% of modern precipitation), with a lowering of the mean annual temperature by some 2.5–4 °C. The climatic deterioration was felt most strongly in the drier central parts of the Alps. Many rock glaciers were reactivated under these cold and dry conditions (Kerschner, 1980). With the onset of the Holocene, the glaciers receded behind their recent limits beginning with a sequence of readvances and retreats (Patzelt, 1995). The last significant ice advance occurred during the Little Ice Age.

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