Field Trip Stops

Stop 1 – Cellon Section

The famous section is located between 1480 and 1560 m on the eastern side of the Cellon mountain, SSW of Kötschach-Mauthen and close to the Austrian/Italian border. It can be reached within a 15 minutes walk from Plöckenpass.

The Ordovician to Lower Devonian part of the Cellon section is best exposed in a narrow gorge cut by avalanches (Fig. 18). Thus, the German name for the section is "Cellonetta Lawinenrinne".

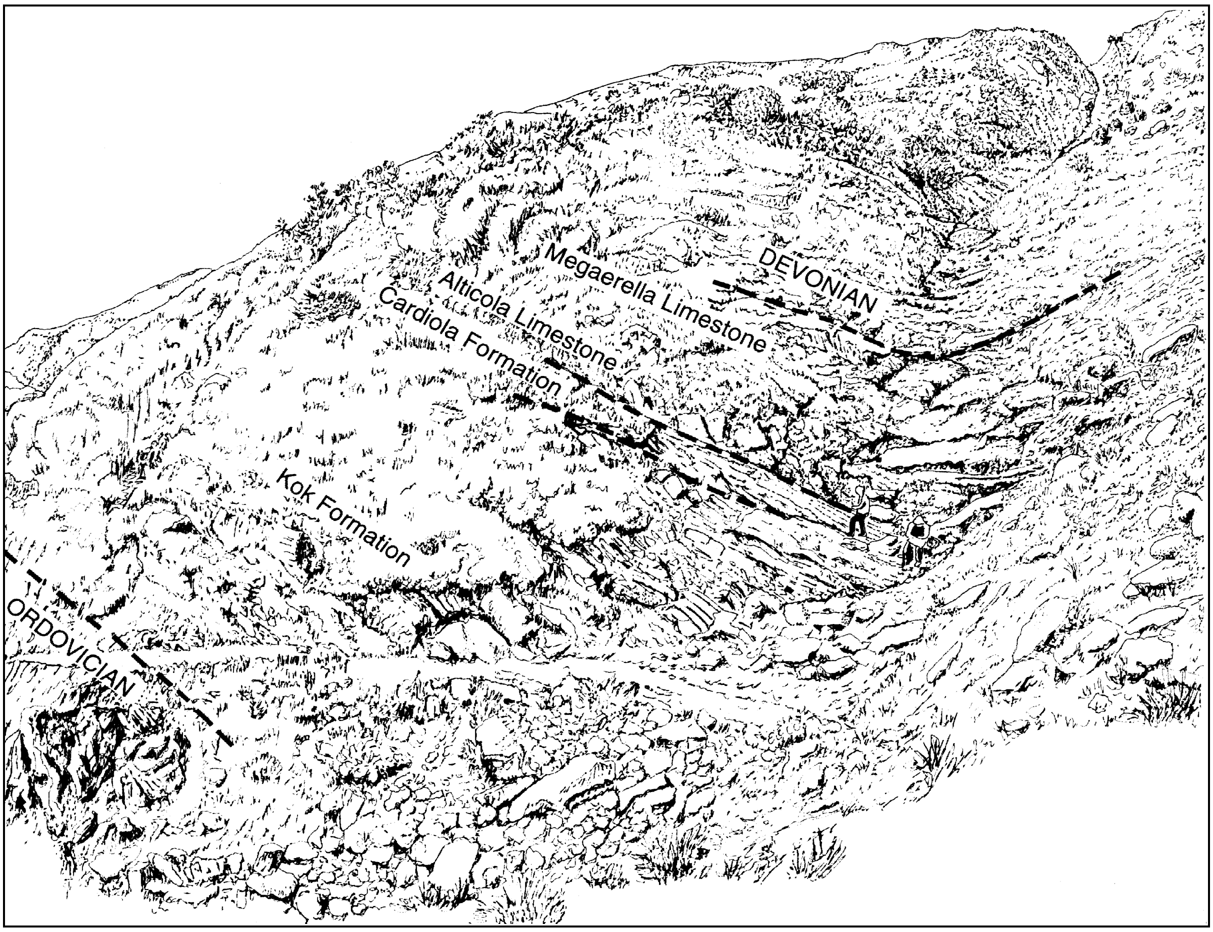

Fig. 18: View of the Cellon section (from HISTON et al., 1999).

Stratigraphy

The Cellon section represents the stratotype for the Silurian of the Eastern and Southern Alps. Nowhere else in the Alps has a comparably good section been found. It has been famous since 1894 when GEYER first described the rock sequence. In 1903 it was presented to the 9th IGC which was held in Vienna. According to v. GAERTNER (1931) who studied the fossils and rocks in great detail, the 60 m thick continuously exposed Upper Ordovician to Lower Devonian section could be subdivided into several formations. Since WALLISER's pioneering study on conodonts in 1964 it still serves as a standard for the worldwide applicable conodont zonation which, however, has been further detailed and partly revised in other areas during the last two decades. Although the conformable sequence, corresponding to the Plöcken Facies, suggests continuity from the Ordovician to the Devonian, in recent years several small gaps in sedimentation have been recognized which reflect eustatic sea-level changes in an overall shallow-water environment. From top to base the following formations can be recognized (Fig. 19):

Lochkovian	80.0 m	Rauchkofel Limestone (dark, platy limestone)
Silurian	8.0 m	Megaerella Limestone (greyish and in part fossiliferous limestone; equivalent to the Pridoli Series)
	20.0 m	Alticola Limestone (grey and pink nautiloid bearing limestone; Ludlow to Pridoli)
	3.5 m	Cardiola Formation (alternating black limestone, marl and shale; Ludlow)
	13.0 m	Kok Formation (brownish ferruginous nautiloid limestone, at the base alternating with shales; Late Llandovery to Wenlock)
Ordovician	5.4 m	Plöcken Formation (calcareous sandstone; Ashgill, Hirnantian Stage)
	6.5 m	Uggwa Limestone (argillaceous limestone grading into greenish siltstone above; Ashgill)
	>50 m	greenish and greyish shales and siltstones (Caradoc to Ashgill)

According to SCHÖNLAUB (1985) the Ordovician/Silurian boundary is drawn between the Plöcken and the Kok Formations, i.e. between sample nos. 8 and 9. In the Plöcken Fm. index fossils of Hirnantian age (brachiopods, trilobites, conodonts) clearly indicate a latest Ordovician age (JAEGER et al., 1975; FERRETTI & SCHÖNLAUB, 2001; SCHÖNLAUB & SHEEHAN, 2003). These strata represent the onset of the end-Ordovician - Lower Silurian transgressive cycle known from many places in the world (SCHÖNLAUB, 1988).

According to conodonts and graptolites from the basal part of the overlying Kok Fm. the equivalence of at least six graptolite and two conodont zones are missing in the Lower Silurian. Renewed sedimentation started in the late Llandovery within the range of the index conodont *P. celloni.*

At present the precise level of the Llandovery/Wenlock boundary can not be drawn. Graptolites and conodonts, however, indicate that this boundary should be placed between levels nos. 11 and 12. Consequently, the rock thickness corresponding to the Llandovery Series does not exceed some three meters.

According to SCHÖNLAUB in KRIZ et al. (1993) the boundary between the Wenlock and the Ludlow Series can be drawn in the shales between sample nos. 15 B1 and 15 B2. Apparently, this level most closely corresponds to the stratotype at quarry Pitch Coppice near Ludlow, England. We thus can assume an overall thickness of some 5 m for Wenlockian sedimentation. By comparison with the Bohemian sections the strata equivalent to the range of *Ozarkodina bohemica* are at Cellon extremely condensed suggesting that during the Homerian Stage sedimentation occurred mainly during the lower part. With regard to the foregoing Sheinwoodian Stage it may be concluded that at its base the corresponding strata are also missing or represented as the thin shaly interval between sample nos. 12 A and 12 C. At this horizon the *M. rigidus* Zone clearly indicates a late Sheinwoodian age.

By correlation with Bohemian sequences and the occurrence of index graptolites for the base of the Pridoli, the Ludlow/Pridoli boundary is drawn a few cm above sample No. 32

(SCHÖNLAUB in KRIZ et al., 1986). This horizon lies some 8 m above the base of the Alticola Lst. The corresponding sediments of the Ludlow have thus a thickness of 16.45 m.

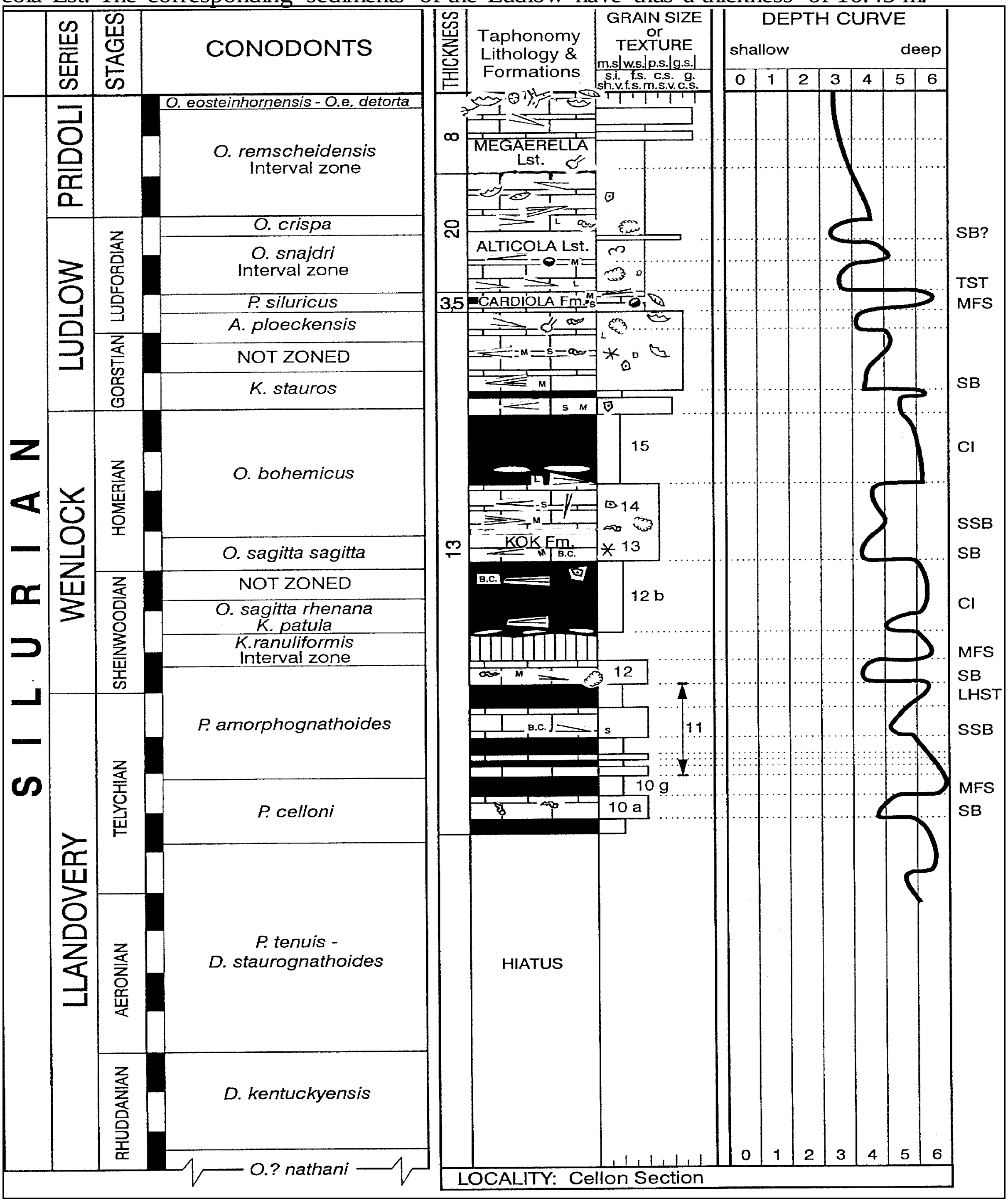

Fig. 19: Conodont stratigraphy, lithology, grain size, significant taphonomic features and depth curve of the Silurian of the Cellon section. Modified from SCHÖNLAUB, 1997.

At Cellon the Silurian/Devonian boundary is placed at the bedding plane between conodont sample nos. 47 A and 47 B at which the first representatives of the index conodont *Icriodus woschmidti* occur. It must be emphasized, however, that the first occurrences of diagnostic graptolites of the Lochkovian is approx. 1.5 m higher in the sequence. JAEGER (1975) recorded the lowermost occurrences of *M. uniformis, M.* cf. *microdon* and *Linograptus posthumus* in sample no. 50. The Pridoli may thus represent a total thickness of some 20 m.

Lithology and Microbiofacies

The first facial investigation at the Cellon section was carried out by FLÜGEL (1965). BANDEL (1972) studied the facies development of the Lower and Middle Devonian in the central part of the Carnic Alps. Middle and Upper Devonian and Lower Carboniferous strata (exposed as steep cliffs and on top of Cellon) were investigated by KREUTZER (1990). Photomicrographs from the Ordovician to Lower Carboniferous sequences comprising the whole Cellon section were published by KREUTZER (1992b) and a preliminary study of the Silurian was given by KREUTZER (KREUTZER & SCHÖNLAUB, 1994). Current work on the cephalopod limestone biofacies in the Carnic Alps with regard to the paleogeographical setting during the Silurian has highlighted many interesting microfacial aspects of the predominantly calcareous sequence.

Uggwa Limestone (Beds 1-4)

This up to 6 m thick limestone horizon comprises indistinctly bedded grey to coloured pelagic "flaser" limestones with ostracod, cystoid and bryozoan debris layers. Sceletal grains consist of brachiopods, ostracods, bryozoans, agglutinate forams, rare cephalopods, trilobites, conodonts and acritarchs.

Plöcken Formation (Beds 6-8)

The 5.40 m thick Plöcken Fm. comprises impure coarse-grained limestones in the lower part consisting of skeletal grains of echinoderms, brachiopods, ostracods, trilobites and conodonts. These grainstones grade into calcareous sandstones and siltstones and shales.

Kok Formation (Beds 9-20) (Fig. 20)

Bed 9: Base of the Silurian sequence: At the level of the Ordovician/Silurian boundary the transition from the greenish silts-shales of the Plöcken Fm. to the carbonate sequence of the Upper Llandovery is marked by the occurrence of flattened nodules approximately 3-5 cm in diameter which appear to be micritic, dark grey-black in colour, quite dense and showing iron weathering: The overlying shales and carbonate layers are badly deteriorated: Fossil content not apparent.

Bed 10: Again a series of shales and thin carbonate beds: level E is the best preserved and shows trace fossil features at the base and the first development of "crust" like shales otherwise fossil content not apparent although a trilobite fauna has been described from this level.

Bed 11: The base is marked by micritic lenses or nodules with "crusts". The overlying shales have a crinoid, trilobite and brachiopod fauna towards the top of the sequence: The first occurrence of nautiloids is at the base of the Wenlock with levels of alternating shales and of reddish-grey micritic carbonate levels which have upper and lower crusts. There is a nautiloid fauna both in the shales and limestones. The shales show flow features around the lenses and the nautiloids are enclosed within the shales. They are small to medium in dimension with an abundance of medium nautiloids towards the top of the sequence. They are parallel to bedding with both body chamber and apexes preserved and have an outer oxidised coating only in the carbonates. A change may be noted up the sequence in that the nautiloids become relatively more abundant in the carbonate levels whereas previously they were more abundant in the shales.

General remarks: Thin beds of ferruginous limestone, sometimes bioturbated, are intercalated at the base of the Kok Formation in dark shales locally rich in small brachiopods. At the top of bed 12, a thin and lenticular calcareous horizon (12b) in shales has provided an important cardiolid fauna (KRIZ, 1999). This is a cephalopod wackestone the matrix of which bears many ostracods, echinoderms, rare small bivalves and gastropods. Many muellerisphaerida are present in darker bituminous micritic areas.

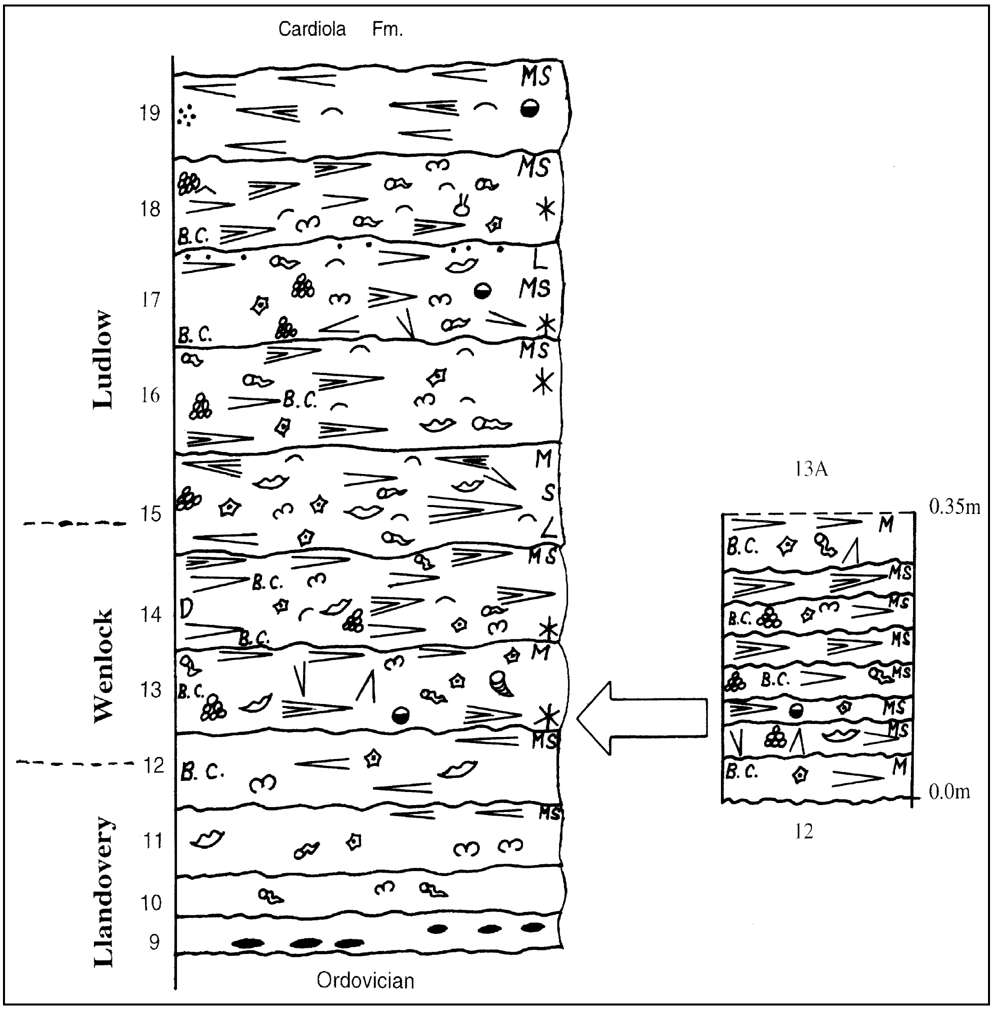

Fig. 20: Taphonomy of the Kok Formation. Note detail of small scale cyclic repetition of beds indicating changes in the hydrodynamic regime.

Starting from bed 13, the limestone becomes thicker and more massive. The reddish colour and the intensive bioturbation are the most typical features of the upper part of the Kok Formation up to level 17. This cephalopod wackestone is locally rich in brachiopods, echinoderm debris, trilobites, gastropods and ostracods. Some organisms, mostly cephalopods, reveal peculiar iron-banded coatings. Dolomitization is frequent. Around level 15 B1 a singular grainstone of well sorted equidimensional bioclasts occurs which strongly resembles the coeval horizon of the Rauchofel Boden section. Abundant small thin-shelled bivalves, preserving the two valves still connected, gastropods, trilobites and isolated echinoderm ossicles have iron-stained shells. Shell in shell structures are common there. Starting from around bed 18 the limestone becomes greyer. Pyrite aggregates in the matrix may be occasionally found.

Cardiola Formation (Beds 21-24)

It is represented by bioclastic shelly layers a few centimeters thick (wackestone-packstone) with a sharp base interbedded with dark shales. At the base of the Cardiola Formation (level 21) bioclastic wackestones rich in cephalopods, trilobites, crinoids and ostracods are intercalated in soft micritic sediments. Scouring traces at the top of the soft sediments, debris grainstone at the base of the overlying horizon with enrichment in iron and manganese oxides would exhibit, according to KREUTZER (1992b), the existence of a Fe-Mn hardground. Millimetric pavements of small brachiopods are present in bed 22. When seen in thin-section, they reveal a cephalopod-ostracod bioclastic packstone with abundant brachiopods, but also associated with graptolites, thin-shelled bivalves and micritized grains. Shelter porosity, common orientation of geopetal structures and telescoping of cephalopods have been observed. Sorting is moderate. These shelly laminae decrease in thickness towards the top of the formation and alternate with thin dark bands rich in organic matter and muellerisphaerida, possible ostracods and recrystallized cephalopods.

Alticola Limestone (Beds 25-39)

The Alticola Limestone (Ludlow-Pridoli in age) is distinctive in that it forms the base of the steep slope of the section. The erosive base of the grey dolomitised massive beds contrast sharply with the black shales of the underlying Cardiola Fm. and this reflects an easily recognizable greyish to reddish limestone formation. It has an overall thickness of 20 m and represents a transgressive carbonate series within more stable pelagic conditions (SCHÖNLAUB, 1997). Grey to dark pink limestones represented mainly by a bioclastic packstone with fine-grained micritic matrix with a variety of bed thickness and frequent stylolites are common in the Ludlow with a dominant nautiloid fauna. The beds decrease in thickness in the Pridoli and alternate with interbedded laminated micrites with a dominant nautiloid and brachiopod fauna. Several deepening events marked by the development of black shales have been documented within the uppermost levels of the Pridoli. Cephalopods are abundant, together with crinoids, trilobites, large gastropods and ostracods. Iron-coatings, mostly around trilobites, are again present. Bioturbation is common.

Megaerella Limestone (Beds 40-47A)

The Megaerella Limestone (Pridoli in age) comprises the upper Pridoli and Silurian/Devonian boundary transgressive sequences of carbonates rich in biodetritus, lenticular micrites and black shales. It has a thickness of 8 m and forms the steep step at the top of the section. Light grey limestones (wackestone to packstones) with cephalopods, ostracods, echinoderm debris and trilobites are dominant. A particular level of juvenile nautiloids (RISTEDT, 1968) occur in bed 40. Bryozoans (*Fenestella* s.l. sp. and a small indeterminate cryptostome [Wyse JACKSON, pers. comm.]) occur on a distinct bedding plane above the Silurian/Devonian boundary together with bivalves. Complete specimens of *Scyphocrinites* (HAUDE, pers. comm.), solitary corals and articulated cridoid stems are common in the lower beds of the Lochkov.

According to KREUTZER (1994) the bathymetric environment for the Upper Ordovician to Devonian sequence can be described as follows:

As early as in the Ordovician a facial differentiation can be recognized for the carbonates. The Cellon section with its Uggwa Limestone development (sample 1-5) represents the late Ordovician Uggwa Facies which is time-equivalent to the Wolayer Limestone of the Himmel-berg

Facies exposed, e.g., at the Rauchkofel-Boden section. Based on conodonts the Uggwa Limestone is well dated as being Ashgillian in age. According to DULLO (1992), the two formations represent the near-shore parautochthonous cystoid facies (Wolayer Limestone) and an off-shore basinal debris facies (Uggwa Limestone), respectively.

At the end of the Ordovician in the Carnic Alps a regression occurred. The Uggwa Limestone bed nos. 1-4 characterized by pelagic faunal elements, are followed by limestones composed of subtidal components of the Plöcken Formation (bed nos. 5-8). A significant unconformity separates the Plöcken Fm. from the overlying Kok Fm.

Transgression of the Kok Formation started in the Cellon section in the Upper Llandovery (bed no. 9). In contrast to the Cellon section the Rauchkofel section located some 8 km to the northwest exhibits a considerably reduced sequence. At Cellon the basal Silurian succession represents a moderately shallow environment which may have lasted until the Llandovery/Wenlock boundary or until the very beginning of the Wenlock. Sample 11 exhibits a very shallow to intertidal environment. During the remaining part of the Wenlock a transgressive trend can be recognized. However, at the Wenlock/Ludlow boundary (bed nos. 15A-F) some strata may be missing reflecting either submersion or reduced sedimentation.

During deposition of the Cardiola Formation (bed nos. 21-24) contemporary non-deposition may have occurred. Black limestone and shale beds with radiolarians alternate with pelagic limestone beds indicating an offshore environment. The following Alticola Limestone (bed nos. 25-39) reflects stable conditions in a pelagic environment which terminated in a regressive pulse (bed no. 40). With the beginning of the Megaerella Limestone (nos. 41-47A) a further transgressive trend can be inferred.

Starting in the Lochkovian Stage (bed 47B and >; Rauchkofel Limestone) and ranging to the Upper gigas Zone of Frasnian age (top region of the Cellon cliff) the Devonian transitional facies represents a fore-reef facies. While this slope facies accumulated at Cellon, only a few kilometers to the palinspastic SSW (today seen at the Kellerwand region) more than 1000 meters of Devonian shallow-water limestones were deposited. Moreover, coeval carbonates of pelagic origin, i.e. pelagic limestone facies of the Rauchkofel Nappe, with a markedly reduced thickness of not more than 100 meters were deposited within short distances to the NNE (SCHÖNLAUB, 1979, 1985; KREUTZER, 1990, 1992a, b).

During the crepida Zone of the Famennian a short-lasting regression occurred. In the Upper Famennian and Lower Carboniferous uniform cephalopod limestones were deposited (Pal and Kronhof Limestone, respectively). At the beginning of the Visean the flysch of the Hochwipfel Formation transgressed upon the Kronhof Limestone and limestone deposition ended.

In more detail the Devonian to Lower Carboniferous succession is subdivided into the following formations (KREUTZER, 1992). It represents the transitional facies between the southwestern shallow-water realm and the eastern to northeastern deep-water settting:

80 m	well-bedded pelagic Rauchkofel Lst.: dark grey and black plate limestones with occasional organodetritic interbeds (Lochkov);
120-150 m	Kellerwand Lst.: well-bedded yellowish tentaculite limestones alternating with skeletal debris layers (Pragian to Lower Emsian);
120 m	Vinz Lst.: well-bedded dark grey platy limestone interbedded with detritic layers (Emsian);
150-200 m	Cellon Lst.: grey massive limestone beds composed of pelagic biogenes, bioclasts and debris layers (Eifelian-Givetian);
50-100 m	Pal Lst.: greyish to reddish and also pinkish cephalopod limestone (Frasnian to Famennian);
1-3 m	Kronhof Lst.: greyish to reddish cephalopod limestone (Tournaisian).

A short distance to the west of the peak of Cellon at the famous Grüne Schneid section the Devonian/Carboniferous boundary beds are excellently exposed. The detailed distribution of conodonts, goniatites and trilobites as well as the lithology and major and trace element content was recently studied by an international working group (see SCHÖNLAUB et al., 1992).

The Ordovician-Silurian Boundary Event
(SCHÖNLAUB & SHEEHAN, 2003)

The mass extinction at the end of the Ordovician led to the disappearance of about 100 families, which represented about 22% of all marine families (SEPKOSKI, 1982, 1993). This demise affected mainly trilobites, brachiopods, echinoderms, stromatoporoids, corals, bryozoans, ostracods, bivalves, cephalopods, graptolites, conodonts, chitinozoa and acritarchs. In comparison with the great dying at the P/T boundary, this was the second largest catastrophe in the Phanerozoic.

During the last years a number of explanations have been suggested for this extinction, such as the species/area-effect as a negative consequence of the global ice age with an associated regression (BRENCHLEY, 1995; HARPER & JIA-YU, 1995; OWEN & ROBERTSON, 1995), increased sedimentation with related pollution due to a significant sea-level decrease (WYATT, 1995), or changes in the composition of sea water and precipitates (ORTH et al., 1986; WILDE et al., 1986; MELCHIN et al., 1991; GOODFELLOW et al., 1992; WANG et al., 1992, 1993; LONG, 1993). Furthermore, in a few areas, such as South China, the Canadian Arctic, on Anticosti Island, and in South Scotland (Dob's Linn stratotype area), elevated Ir contents were found in boundary layers. However, they have been interpreted to be of terrestrial origin (WANG et al., 1995).

The global ice age at the end of the Ordovician is unusual because it occurred at a time when the atmosphere had increased CO_2 contents and, thus, the Earth should have had a relatively stable greenhouse climate (BRENCHLEY et al., 1994). At that time the atmospheric CO_2 content was supposedly 14 to 16 times higher than today (BERNER, 1990, 1992, 1994; CROWLEY & BAUM, 1991; GRAHAM et al., 1995; MORA et al., 1996).

New studies seem to have clarified the long-standing question regarding the exact timing of the end-Ordovician mass extinction. According to SUTCLIFFE et al. (2001) and SHEEHAN (2001) this event occurred at the base of the Hirnantian Stage of the Upper Ordovician with the beginning of the graptolite Zone of *Normalograptus extraordinarius*. At that time the ratios of the stable isotopes of C, O and S changed, which was explained as a result of biomass reduction, temperature increase, and a short-term flooding of the continental shelf with anoxic water (GOODFELLOW et al., 1992). The latter supposedly was the ultimate cause of the mass extinction. However, an opposite trend was observed for the Hirnantian Stage in middle Sweden by MARSHALL & MIDDLETON (1990).

Based on recent stratigraphic and geochemical studies, the following scenario has been proposed for the Upper Ordovician: Beginning with the so-called gracilis-transgression of the older Caradoc Series, the global sea level increased until the end of the Ordovician (ROSS & ROSS, 1992). This second-order cycle was, however, overprinted by a regressive-transgressive trend in the late Ashgill Series lasting for about 0.5 to 1 million years, i.e. during the early to middle Hirnantian Stage. This trend was caused by glacio-eustatic changes and led to a sea level dropping of about 100 meters (BRENCHLEY & NEWELL, 1980; SHEEHAN, 1988, 2001; BRENCHLEY & MARSHALL, 1999; BRENCHLEY et al., 1991, 1994). The glaciation extended over an area of approx. 30 million square kilometers of the southern hemisphere. It resulted in a significant decrease in the average temperature, as well as in an increased productivity due to increased oceanic circulation. The latter is evident from anomalies in the oxygen and carbon isotope ratios of marine carbonates and in the organic carbon reservoir. This deterioration of the climate and the overall changes of the nutrient supply was held responsible for the first pulse of the twofold mass extinction. After the faunal demise the opportunistic Hirnantia Fauna started to invade middle and higher latitudes. This fauna reached its highest diversity within the *N. extraordinarius* Zone.

At the base of the graptolite Zone of *N. persculptus*, both isotope curves show an opposite trend (MARSHALL et al., 1994; BRENCHLEY et al., 1994; WANG et al., 1994). This signal is assumed to indicate the end of the glacial period, as well as a decreased production of biomass. This condition seems to have been the reason for a second mass extinction shortly before the end of the Ordovician.

Facies sequence data from Upper Ordovician deposits in the Carnic Alps also reflect a regressive-transgressive sedimentary pattern during the Hirnantian Stage (SCHÖNLAUB, 1988; SCHÖNLAUB & SHEEHAN, 2003) (Fig. 21). The regressive trend is well documented in the upper part of the Uggwa Limestone and culminates in the bioclastic Plöcken Formation of the late Hirnantian Stage. Newly recovered conodonts and graptolites from this formation seem to indicate a level immediately below the Ordovician/Silurian boundary (FERRETTI & SCHÖNLAUB, 2000; SCHÖNLAUB & SHEEHAN, 2003).

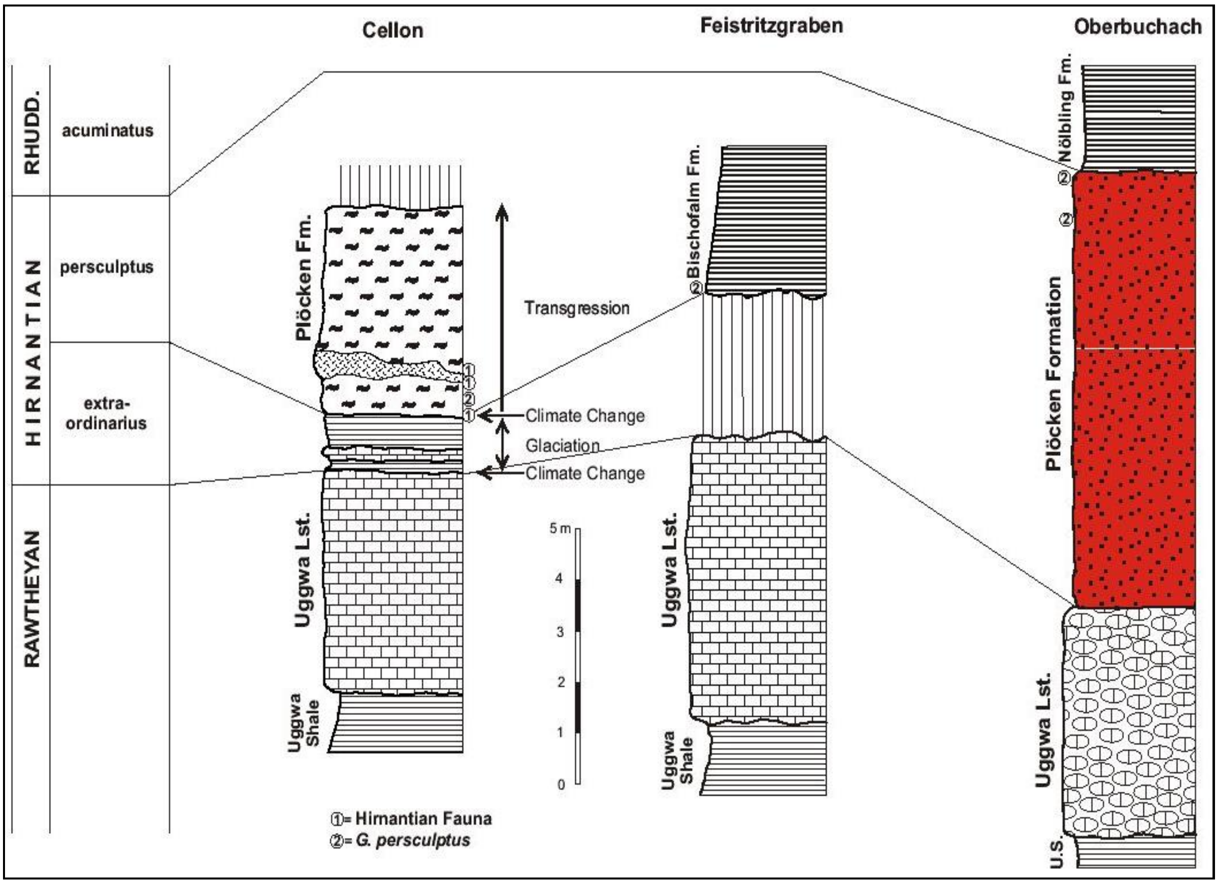

Fig. 21: The Upper Ordovician succession of the sections Cellon, Feistritz Gorge and Oberbuchach (Carnic Alps and Western Karawanken Alps) with inferred glacio-eustatic events.

In the Carnic Alps clastic sediments are the dominating lithologies during the Caradocian and early Ashgillian Series. They are succeeded by fossiliferous limestones known as the up to 20 m thick bryozoan and cystoid bearing Wolayer Lst. and the coeval 4-6 m thick Uggwa Lst., respectively. The latter represents a slightly deeper open marine setting being formed during the Rawtheyum Stage at the beginning of the upper Ashgillian.

The Uggwa Lst. is overlain by greyish and greenish siltsones with thin interbedded limestone layers. The transition from calcareous to pelitic sedimentation occurs approx. 0.40 m below the first appearance of the Hirnantia fauna characterizing the base of the succeeding Plöcken Formation. This fauna is associated with trilobites, e.g., *Mucronaspis m. mucronata* and graptolites such as *Normalograptus persculptus*.

The change from the Uggwa Lst. to the siltstones reflects significant environmental changes at the beginning of the Hirnantian Stage (Fig. 22). Hence, the greenish pelitic shales atop the Uggwa Lst. may represent glacio-marine sediments in distal regions of the icesheet covering large parts of southern Gondwana. They are separated from the overlying transgressive Plöcken Fm. by a distinct unconformity.

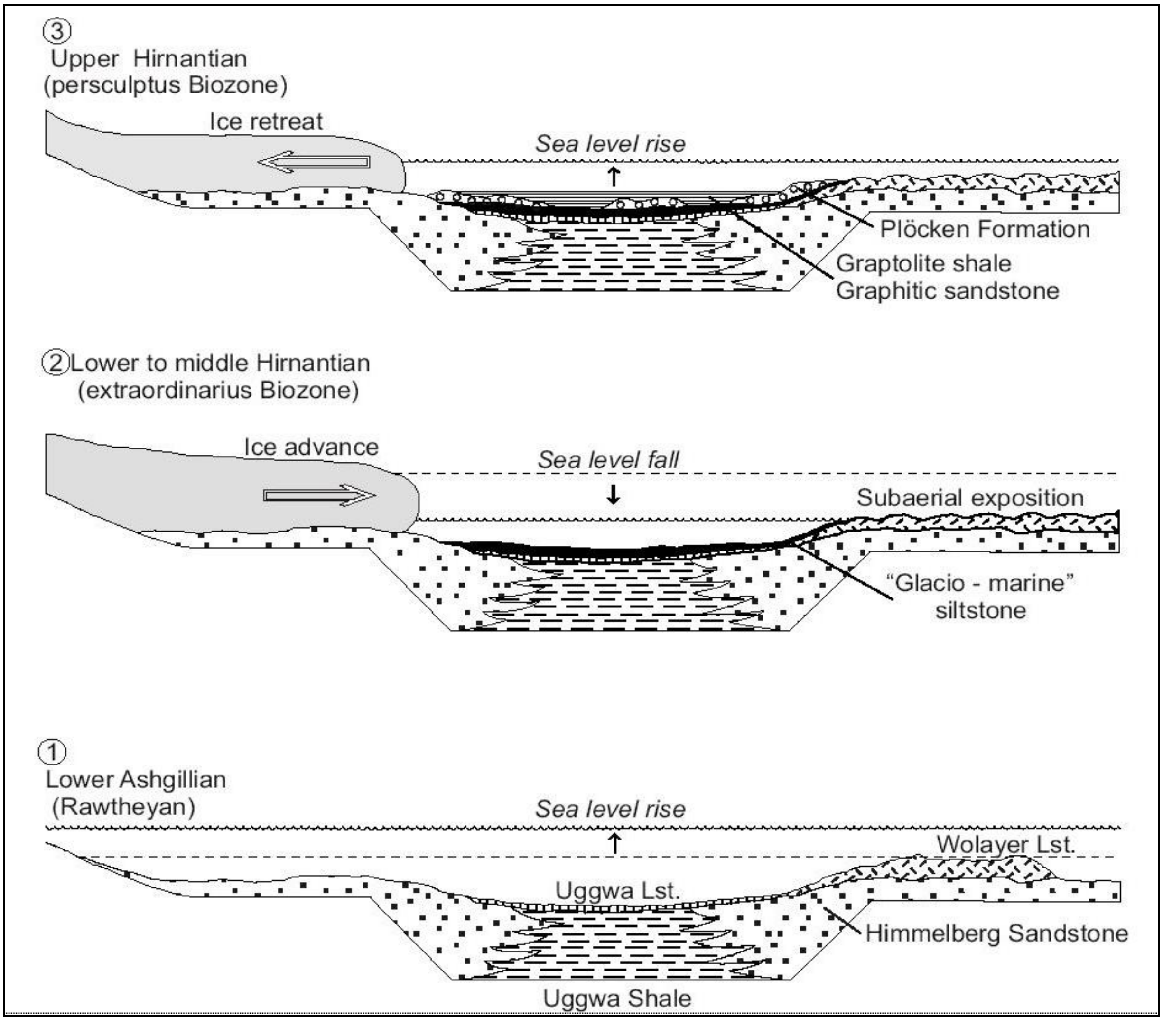

Fig. 22: Model showing the governing sedimentation pattern during the Ashgillian Series in the southern Alps.

At the Cellon section the Plöcken Fm. attains a thickness of 5.40 m. The lower 0.80 m thick portion is composed of arenaceous siltstones followed by impure limestones and calcareous sandstones with layers of bio- and lithoclasts. The fossil debris mainly represents disarticulated brachiopod shells but also bryozoans, trilobites, ostracods and conodonts are quite abundant. The whole package is strongly bioturbated, partly graded and convolute bedding and channeling occurs. This lithology suggests a storm-dominated shallow water environment which formed during the retreat of the ice in the *N. persculptus* Biozone.

Representatives of *N. persculptus* occur approx. 0.25 m above the base of the Plöcken Fm. Hence, this index graptolite testifies the upper Hirnantian Stage during which the transgression started on a global scale. Due to local tectonic uplifts, however, in the Carnic Alps a gap in sedimentation occurred at the base of the Silurian. Thus, at the Cellon section the equivalences of the lower and middle Llandovery are missing. Continued sedimentation across the passage from the Ordovician to the Silurian seems to have only occurred in the basinal black shale environment of the Bischofalm Facies.

Stop 2 – Devonian Succession at Mount Freikofel

Mt. Freikofel is located to the east of the Plöckenpass (Passo di Monte Croce Carnico) and can be reached by following Trail # 403 from the Plöcken Haus (1215 m) to the trail head of trail # 401 which climbs to the top of Mt. Freikofel (Cuelat, 1757 m). The trail on the Austrian side shows good exposures of the Frasnian/Famennian succession whereas the branch on the Italian side follows an old army track and shows best exposures of the Lochkovian to Middle Devonian succession.

Note: The military track is not difficult to walk (it was made for mules) but it is not secured and drops off steeply to the sides. It is not recommended for those afraid of heights and great care must be taken not to dislodge stones. Good foot wear (boots) is essential.

Mount Freikofel exhibits a spectacular section which spans almost the entire Devonian, it is easily accessible, well preserved and well exposed.

Based on lithological criteria the succession can be subdivided into five units (Fig. 23):

Unit 1 Basal dark grey platy and lumpy limestones (~77 m thick).
Unit 2 Yellow-grey lumpy to nodular bedded limestones with intercalated calcarenite beds (~74 m thick).
Unit 3 Massive lithoclastic limestone with reefal debris and lithoclasts (~68 m thick).
Unit 4 Bedded lithoclastic limestone with intercalated clacarenite units, increasing bed thickness up section (~56 m thick).
Unit 5 Grey stylo-bedded fine-grained limestone and grey to pink burrow-mottled limestone with intercalated calcarenite and calcirudite beds (~36 m thick).