Annales de la Société Géologique de Belgique, T. 113 (fascicule 2) - 1990, pp. 53-73 Février 1991

THE STAVELOT MASSIF FROM CAMBRIAN TO RECENT A SURVEY OF THE PRESENT STATE OF KNOWLEDGE

by

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(17 figures)

RESUME.- A l'occasion de la réunion des géologues Meuse-Rhin au Mont Rigi (Hautes Fagnes, Massif de Stavelot), les 18-19 mai 1990, une revue de la géologie du Massif de Stavelot a été préparée. Elle illustre l'état actuel des connaissances, des recherches géologiques pendant la dernière décade et les problèmes qui subsistent.

ABSTRACT.- At the occasion of the annual Meuse-Rhine geologists meeting at Mont Rigi (Hautes Fagnes, Stavelot Massif) on May 18-19, 1990, a review has been prepared on the geology of the Stavelot Massif. It illustrates the actual state of the art, achievements and remaining problems, of the geological investigations during the past decade.

INTRODUCTION

The Stavelot Massif (also referred to as Stavelot-Venn Massif) consists of Cambro-Ordovician rocks (fig. 1). It is one of the four areas in the Ardenne where Cambrian rocks occur at or near the surface. However it is distinguished from the other three areas (Rocroi, Serpont and Givonne) by the presence of unequivocally Ordovician rocks, which form a sequence of at least 900 m in thickness, representing the Early (and perhaps the base of the Middle) Ordovician. Deposits of the Middle and Late Ordovician, the entire Silurian and perhaps even the Early Lochkovian are missing in the stratigraphic record of this region. Apparently, sedimentation started again during the Late Lochkovian as deposits of this age fringe the border of the Stavelot Massif. In this context, it should be noted, that the «Grès de Petites Tailles» (Baraque Fraiture, southernmost part of Stavelot Massif) has been interpreted as a fluviatile «pre-Devonian» (Silurian?) sediment, deposited before the main Lochkovian transgression (Geukens, 1986).

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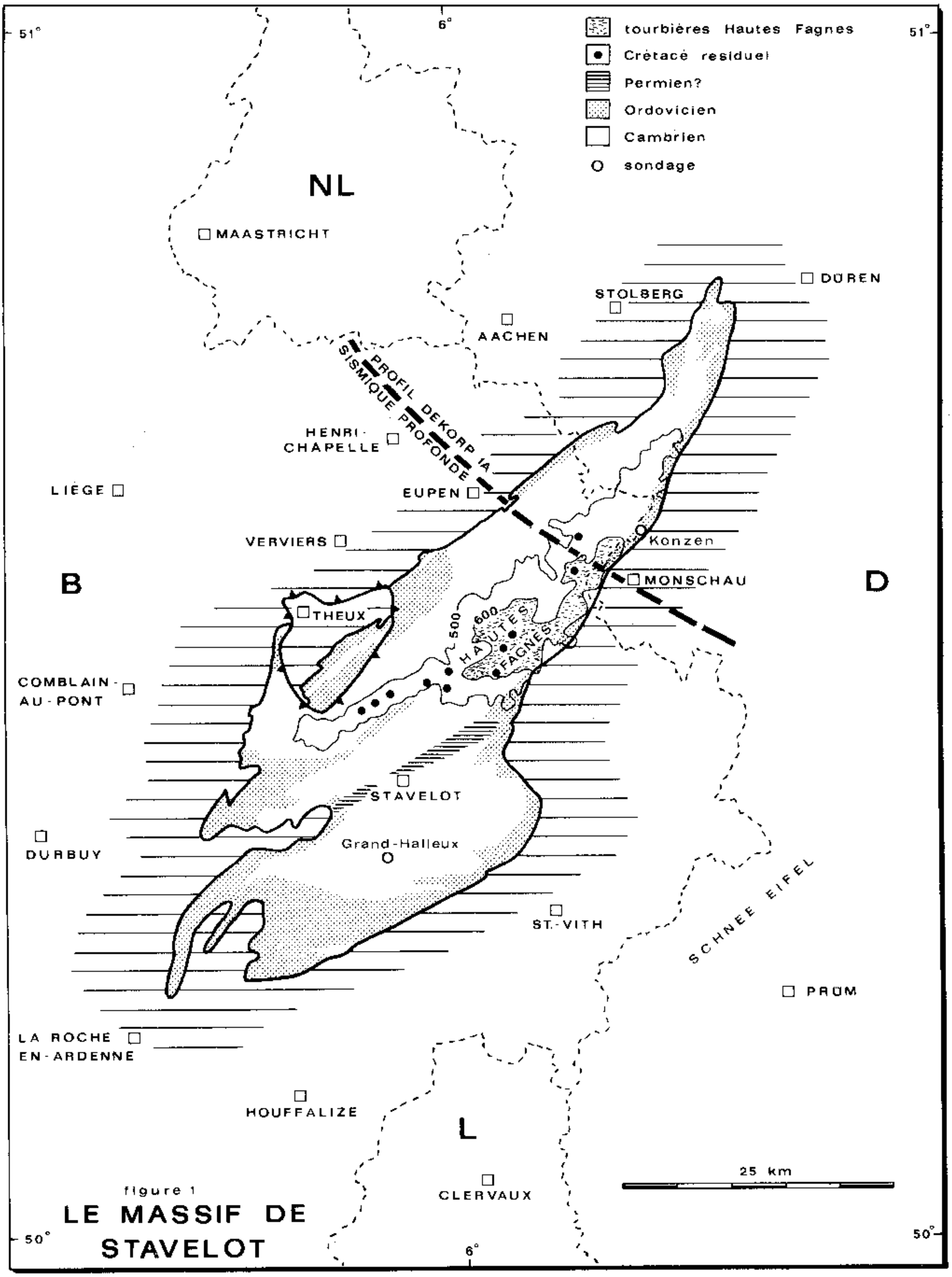


Figure 1.- Outline of the Stavelot Massif and the tectonic window of Theux. The 500 m and 600 m contour lines are only shown in the northern part of the massif, where locally residual Late Cretaceous deposits occur. However, it should be noted that several parts of the southern Stavelot Massif are also above 500 and 600 m altitude.

Numerous bio- and lithostratigraphical, petrological, sedimentological, structural, geophysical and morphological studies have been carried out in this area. Field observations were complemented by borehole information, e.g. at Grand Halleux (see Vanguestaine, 1978) in the south and at Konzen (Walter & Wohlenberg, 1985) in the north. Additionally seismic reflection investigations in the northeastern part of this area have revealed the structural style of the Caledonian and Variscan fault pattern «down to the Moho discontinuity» (Betz *et al.*, 1988). In spite of all this information, many questions remain. The present review of achievements and problems will illustrate this.

STRUCTURAL STYLE

(R.W., J.B., L.D., F.Ge., L.H.)

The Stavelot Massif forms the northeast end of the Ardenne Anticlinorium, which further to the SW also includes the Serpont uplift and the Rocroi Massif. The oldest outcropping rocks are Early Cambrian Deville deposits, which are overlain by the Middle-Late Cambrian Revin strata, and on top of which rest in the Stavelot Massif the Early-Middle Ordovician Salm sediments. The Caledonian tectonic movements must have already shaped the final outline or the Ardenne Anticlinorium. However, the E-W trending Caledonian folds in the Stavelot Massif are overprinted by SW-NE striking Variscan tectonics. Two complexly folded and faulted anticlinal structures can be distinguished in the Stavelot Massif which are separated by the «Graben de Malmédy» with an infill of presumably Permian conglomerates («Poudingue de Malmédy»; Geukens, 1957).

The northwest flank of the Stavelot Massif is characterized by flat-flying thrust faults. One of these is very clearly exposed at Heusy along the expressway Verviers-Prüm (Hance *et al.*, 1989). The famous «Fenêtre de Theux» has been looked upon as a textbook example of a tectonic window ever since Fourmarier's publications (1901-1969). However, this structure has been interpreted in quite different ways (e.g. Geukens, 1959; Graulich, 1984; Klein, 1977a, 1977b; Michot, 1988). Especially the nature of its southern closure was and still is the subject of continued discussion (Robaszynski & Dupuis, 1983).

Four deep seismic reflexion lines have been carried out across the northeast Stavelot Massif and its northern foreland (Meissner *et al.*, 1980, 1981; Durst, 1985; Betz *et al.*, 1988; Dekorp Research Group, in press). These surveys yield information about the deep structures of this area. The line-drawing of the more recent profile Dekorp 1A has tentatively been interpreted on figure 2. A pronounced reflector at a depth of about 3-4 km can be followed over a long distance on each profile. This reflector is interpreted as the southward subsurface continuation of the Midi-Aachen overthrust, but we can only speculate as far as the age and nature of the rocks below this reflector are concerned. Some possible alternative solutions have been forwarded by Walter & Wohlenberg, (1985). Only a deep borehole may produce the final answer. However, north of the «Fenêtre de Theux», the subsurface trace of the Midi-Aachen overthrust is not always very clear, perhaps because it occurs within Caledonian rocks (Dejonghe *et al.*, 1989).

Magmatic rocks are found in the Revinian. Most of them are either interbedded and of rhyolitic/rhyodacitic composition or intrusive and of basaltic composition. There are two main exceptions: the Helle and Lammersdorf sill-type intrusions (referred to as «tonalites») which are of granodioritic to dioritic composition. U-Pb zircon dating of the Helle tonalite has given 381 ± 16 Ma (Kramm & Buhl, 1985). Porphyry Cu-Mo type mineralisation is associated to the Helle intrusion (Van Wambeke, 1956; Weis *et al.*, 1980).

CAMBRO-ORDOVICIAN

(M.V., R.W.)

The paleogeographical evolution of the Stavelot region during the Cambro-Ordovician is marked by important changes in the depositional environment related to alternating uplift and subsidence, sometimes accompanied by active volcanism (Walter, 1980). It is because of the many fault contacts that the lithostratigraphic succession of the Cambro-Ordovician and its thickness are incompletely known. The biostratigraphic zonation based on acritarchs (fig. 3) is still not fully established (cf. Vanguestaine & Van Looy, 1983; Vanguestaine, 1985, 1986).

The presence of occasionally rich and diverse acritarch assemblages in the Cambro-Ordovician strata of the Stavelot Massif illustrates the very low-grade Caledonian metamorphism. Destruction of the organic-walled fossils took only place along the southeast flank of the massif (including the basal Devonian) during the Variscan metamorphism.

The depositional history of the Stavelot Massif during the Cambro-Ordovician was summarized by Von Hoegen *et al.* (1985). Presumably, the Deville quartzites, sandstones and (silty) shales were deposited in a nearshore - gradually deepening - shelf environment (fig. 4). The black shales of the Revin succession reflect conditions of an outer shelf or even deeper basin during the

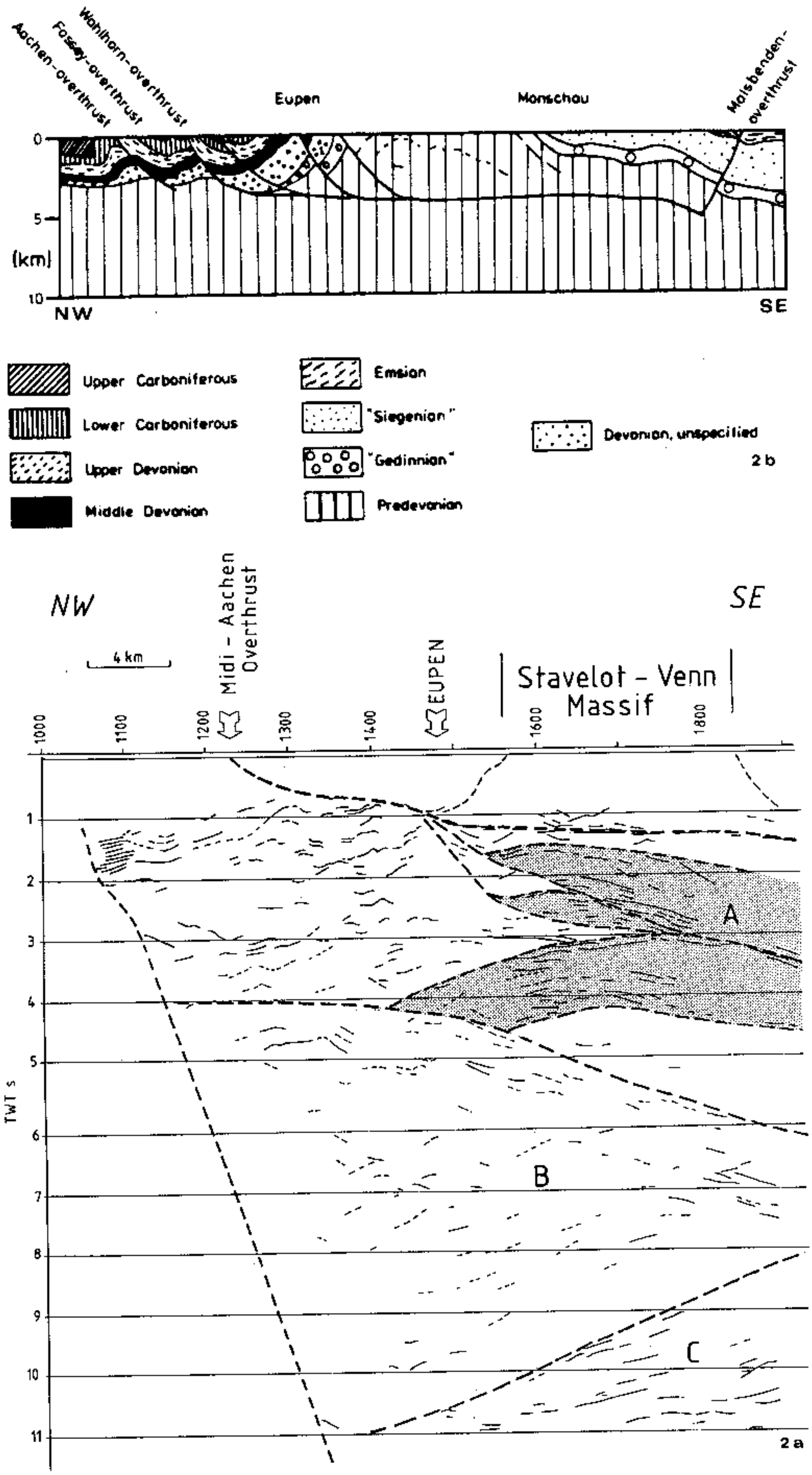


Figure 2.- Geological section through the northern part of the Stavelot Massif (2b, R.W.) and line drawing of the northern part of the seismic reflection profile DEKORP 1-A (2a; J.B., L.H.). Interpretation of figure 2a : A. Duplex structure?; B. Southerly wedge of the Brabant Massif; C. Lower laminated crust.

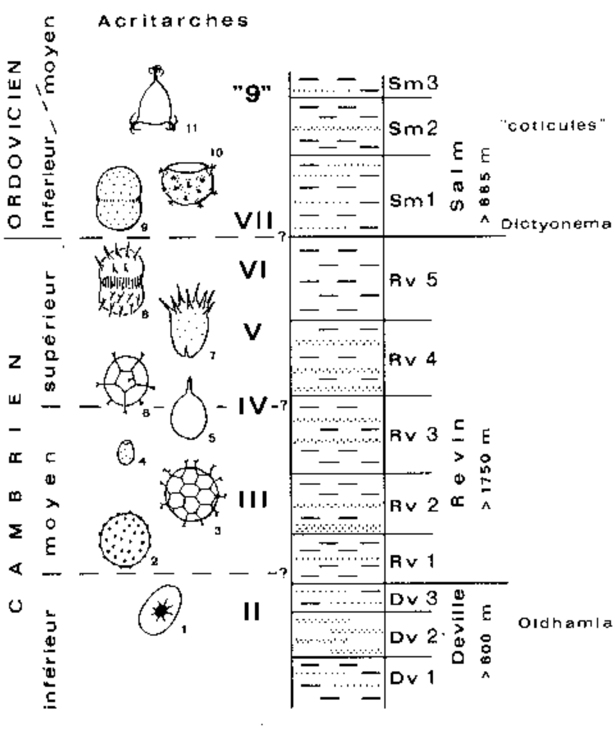


Figure 3.- Synoptic stratigraphy of the Cambro-Ordovician in the Stavelot Massif. Lithostratigraphy after Geukens (*in:* Robaszynski & Dupuis, 1983), chrono-/biostratigraphy after Vanguestaine (*in:* Robaszynski & Dupuis, 1983).

1-11. selected acritarchs: 1. *Archaeodiscina umbonulata;* 2. *Lophosphaeridium tentativum,* 3. *Timofeevia phosphoritica;* 4. *Revinotesta microspinosa;* 5. *Alliumella baltica;* 6. *Timofeevia pentagonalis;* 7. *Trunculumarium revinium;* 8. *Acanthodiacrodium timofeevii;* 9. *Acanthodiacrodium angustum;* 10. *Stelliferidium trifidum;* 11. *Frankea sartbernardensis.* II-VII, «9»: acritarch zones.

Middle and Late Cambrian. Analysis of the heavy mineral assemblages from the intercalated psammitic quartzites has shown that these are reworked from the Deville Formation, which may have been exposed in nearby shoals. The Ordovician Salm deposits represent a prograding delta facies in a regressive environment. Important changes in the petrographic composition of the sediments at the Revin-Salm boundary suggest different sources for the Revin and Salm sediments.

LOCHKOVIAN TRANSGRESSION

(Ph.S., M.S.)

Following an important gap in the stratigraphical record (Middle-Late Ordovician, Silurian), the deposition in the Stavelot area started again during Lochkovian times. Detailed palynological investigations revealed that sedimentation along the northwest flank of the Stavelot Massif (Gileppe, Spa, Nonceveux) did not start until the Late Lochkovian (spore zones M or Si), whereas the oldest Devonian strata along the flanks of the Rocroi Massif to the southwest are of Early Lochkovian age (spore zones N and R; Steemans, 1989a).

The diachronic onset of the Devonian sedimentation is also illustrated by the different age of the basal Devonian strata in the Gileppe section on the northwest flank of the Stavelot Massif (spore zone M) and the Huy section between Namur and Liège (spore zone Si; Steemans 1989a). Even the

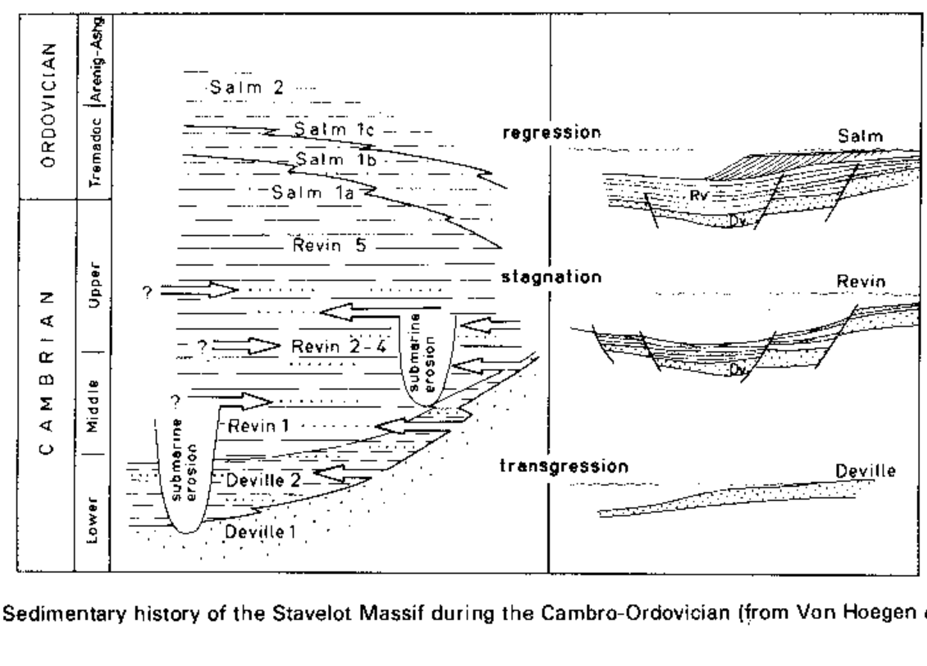


Figure 4.- Sedimentary history of the Stavelot Massif during the Cambro-Ordovician (from Von Hoegen *et al.,* 1985)

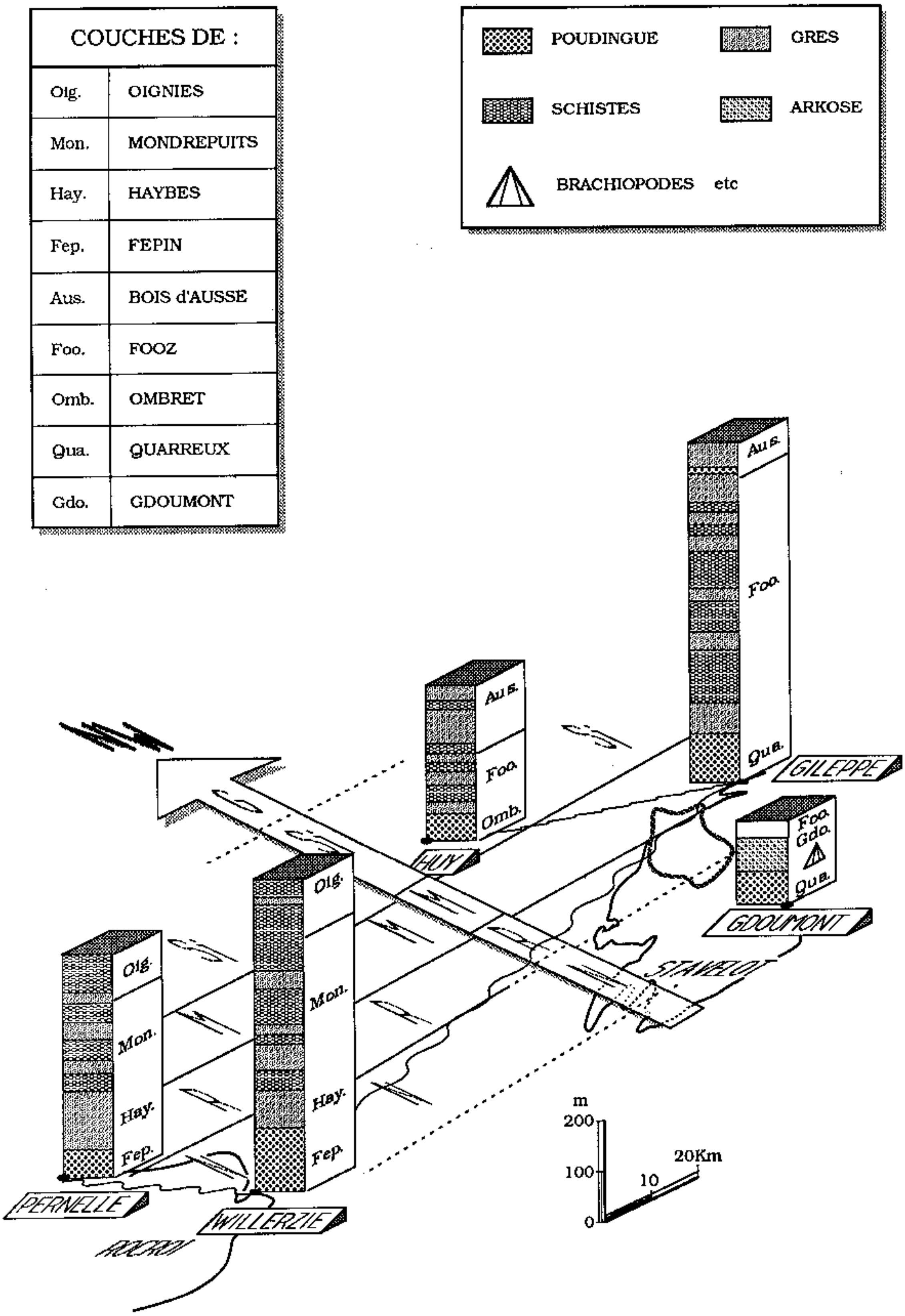


Figure 5.- Diachronic onset of Lochkovian transgression in Belgium as determined by spore assemblages (modified after Steemans, 1989a). The exact age of the Gdoumont section could not be established because of the high coalification rank in that area.

nearby sections of Willerzie (east flank of the Rocroi Massif; spore zone N) and Pernelle (northwest flank of Rocroi Massif; spore zone R) show a marked difference in age of the basal Devonian strata (fig. 5).

These observations serve to emphasize the northwesterly direction of the Lochkovian transgression. This implies that the «Gedinnian» brachiopod assemblages of Gdoumont (SE of Stavelot Massif) may be older (Late Silurian or Early Lochkovian) than the basal Devonian strata along the northwest flank of the Stavelot Massif (cf. Boucot, 1960; Carls, 1977; Steemans, 1989b). Graulich (1951) and Neumann-Mahlkau (1970) also assumed the Lochkovian sea to have transgressed upon the Stavelot area from the southeast. Therefore, it is possible that the southeast flank of the Stavelot Massif was already flooded during the Early Lochkovian (spore zones N or R). This assumption, however, should be tested by means of future investigations. The occasionally extremely high coalification of the basal Devonian strata along the southeast flank naturally poses a problem (see Kramm *et al.*, 1985).

DEVONO-CARBONIFEROUS

(M.J.M.B., J.B., M.S.)

The Stavelot Massif is distinguished from its surroundings by the absence of Devono-Carboniferous sediments, which must have been removed here before the deposition of the «Poudingue de Malmédy» during the Permian (cf. Renier, 1902). Therefore, only indirect evidence is available as to its paleogeographic position during the Devonian

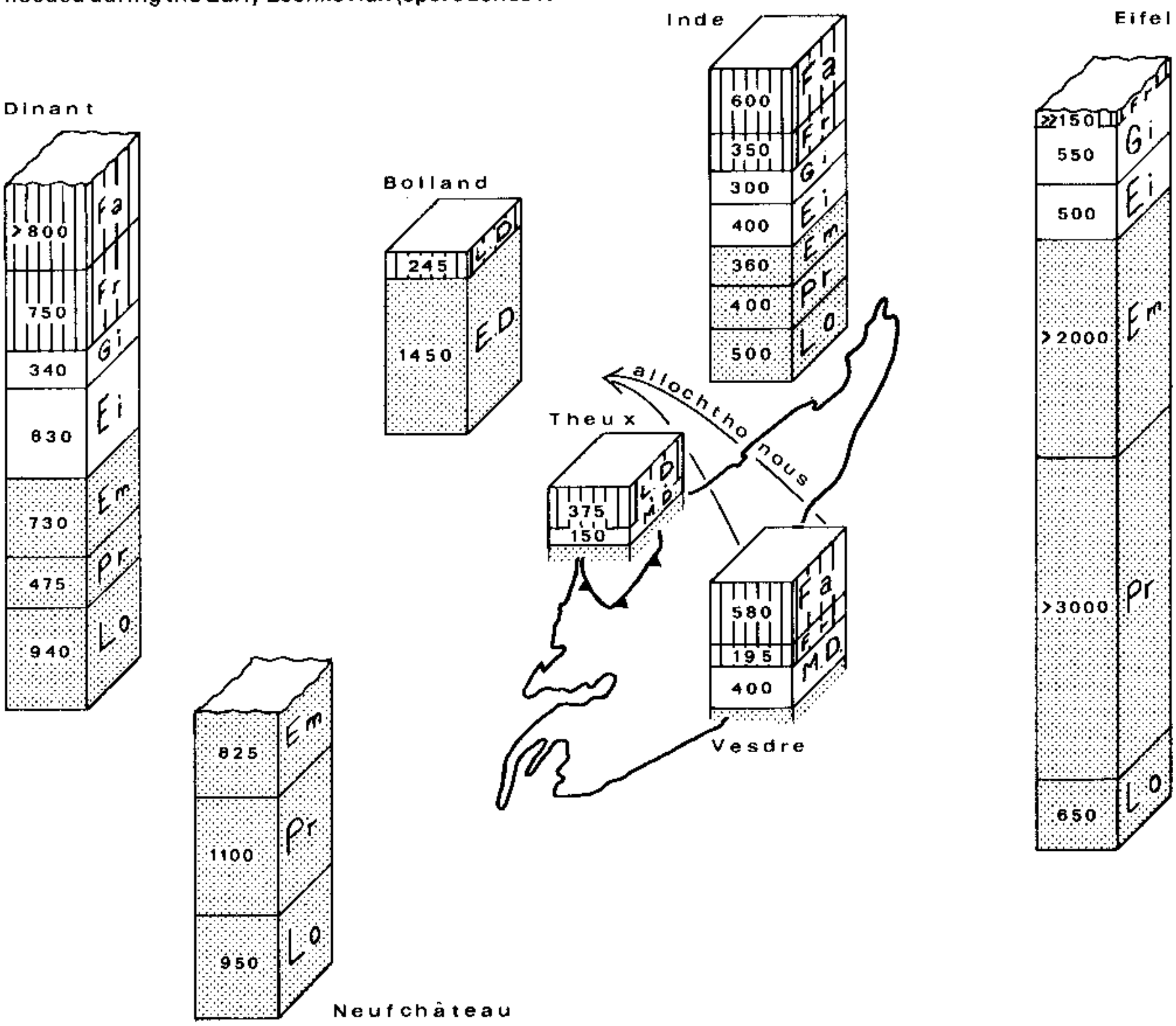


Figure 6.- Thickness variations of Devonian deposits around the Stavelot Massif. After data from Lecompte (1967), Godefroid (1982), Struve (1982), Graulich (1975; and *in* Robaszynski & Dupuis, 1983), Richter (1985) and Walter *et al.* (1985). Note that the column of the allochthonous Vesdre area has been restored to its original relative position in order to obtain a better understanding of the gradual decrease in thickness of the Middle-Late Devonian deposits towards the NW between the Eifel area and the Bolland borehole. The thickness of the Early Devonian in the Theux and Vesdre column is not indicated !

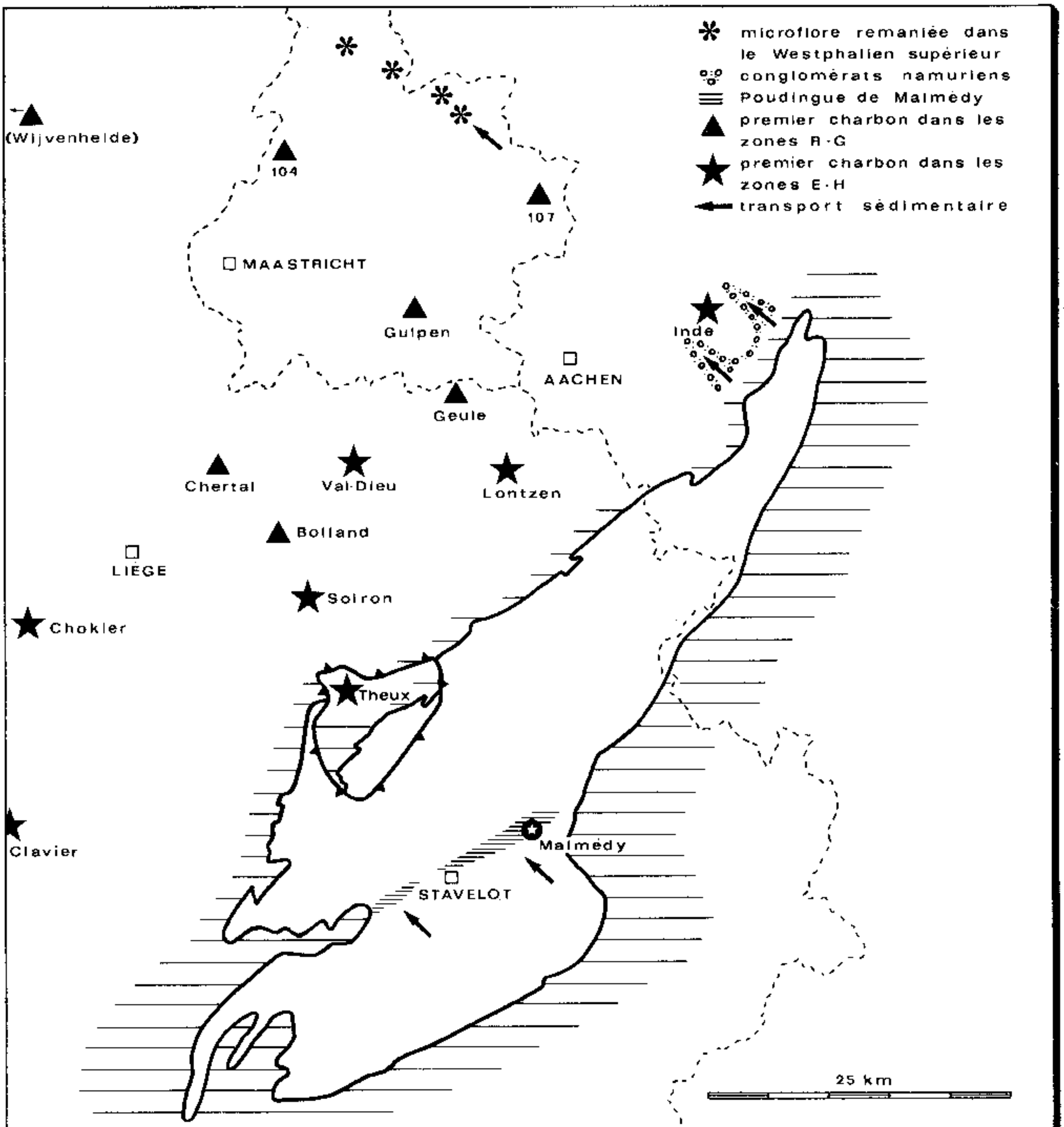


figure 7

SOULÈVEMENT VARISQUE ET ÉROSION

and Carboniferous. Marked differences in the thickness of Devonian deposits to the NW and SE of this massif provide evidence of the variations in the relative rate of subsidence (fig. 6).

During the Early Devonian, the Stavelot Massif separated an area of extremely rapid warping in the SE (5-6000 m in Neufchâteau-Eifel) from an area with a more moderate subsidence in the NW (1250-1450 m in Inde-Bolland). The reduced thickness of the Early Devonian in the «Fenêtre de Theux» suggests that the Stavelot Massif at least in part belonged to the more slowly subsiding Bolland-Inde area.

The pronounced difference in the thickness of the Middle and Late Devonian of the allochthonous Vesdre Massif (originally situated somewhere on the Stavelot Massif south of the «Fenêtre de Theux»; cf. Michot, 1988) and that of the «Fenêtre de Theux» illustrates the differential subsidence within the Stavelot Massif during that time. Notably the Middle Devonian succession is marked by frequent gaps and littoral to perhaps continental deposits both in the «Fenêtre de Theux» and the Vesdre Massif (e.g. Goé macroflora; see Leclercq, 1940; Leclercq & Andrews, 1960; Leclercq & Bonamo, 1971; Lele & Streel, 1969; Lessuise & Fairon-Demaret, 1980). The differential warping during the Middle-Late Devonian may be linked to the Acadian(?) intrusion of the Helle and Lammersdorf tonalites (381 MA; Kramm & Buhl, 1985).

Data on the Carboniferous are limited for the northwest flank of the Stavelot Massif. The presence of reworked Dinantian rocks in the Namurian conglomerates of the Inde area and in the presumably Permian «Poudingue de Malmédy» evidences that Dinantian sediments were also deposited in the Eifel-Neufchâteau area. Namurian conglomerates derived from the south in the Inde area point to Early Namurian uplift and erosion of the northeasternmost part of the Stavelot Massif. However, the onset of the main uplift and subsequent erosion of this area may be correlated with the Middle Westphalian metamorphic event along its southeast flanks (308-312 MA; Kramm *et al.*, 1985a).

VARISCAN UPLIFT AND EROSION

(M.J.M.B., J.B., A.O., A.S., M.S., G.V.)

The end of the subsidence (or perhaps the onset of uplift) in the area of the Stavelot Massif is illustrated by the transition from fully marine to more paralic conditions (indicated by the appearance of coal seams and rootlet horizons). This transition started much earlier along the northwest border of the massif (Early Namurian goniatite zones E-H) than in the area slightly further to the northwest (Late Namurian to earliest Westphalian goniatite zones R-G; cf. Bless & Paproth, 1989).

Uplift and erosion in the Ardenno-Rhenish area started as early as the Late Carboniferous. This is illustrated by the Namurian conglomerates in the Inde area (Burgholz/Walhorn; Gedau/Ardenne) with reworked Dinantian cherts (cf. Hahne & Seidel, 1936), and also by the occurrence of reworked palynomorphs in the Late Westphalian of southern Limburg and the Campine area, which were partly derived from Devono-Dinantian rocks of the Eifel and Rhenish Massif (cf. Bless & Streel, 1976; Van de Laar & Fermont, 1989).

Denudation of the Cambro-Ordovician basement of the Stavelot Massif was completed before the (final) erosion of Devono-Dinantian rocks in the Eifel and Taunus. This is illustrated by the «Poudingue de Malmédy» (fig. 7). This conglomerate fills the WSW-ENE directed «Graben de Malmédy»; which separates the northern and southern parts of the Stavelot Massif (Geukens, 1957). The pebbles of this conglomerate have been reworked from the Cambro-Ordovician basement of the area, as well as from the Early and Middle Devonian of the Eifel and Taunus (Dumont, 1832; Renier, 1902, Maillieux, 1931). The youngest rocks represented in the conglomerate are of Early Dinantian (Ivorian) age as illustrated by the study of conodonts (Smolderen, 1987). The age of the deposit is unknown. Its attribution to the Permian (Renier, 1902) is based only on lithostratigraphic arguments. Paleomagnetic studies indicate that the magnetic pole of the «Poudingue de Malmédy» matches the position of the pole during the Permian (De Magnée & Nairn, 1962; Ozer, 1982).

CALEDONIAN AND VARISCAN METAMORPHISM

(U.K., G.S., M.W.)

Two low-grade metamorphic phases are distinguished in the Stavelot Massif. The first very low-grade one of Late Caledonian age (418 MA; Michot *et al.*, 1973) is characterized by a chlorite-muscovite mineral assemblage in the Deville phyllites of the Grand Halleux area in the southern part of the massif. The radiometric dating of this event was subsequently questioned by Kramm *et al.* (1985a).

The second phase of Variscan age (308-312 MA; Kramm *et al.*, 1985a) affected both the Ordovician (Salm) and Lochkovian rocks along the southern and eastern flanks of the Stavelot Massif (fig. 8; Kramm *et al.*, 1985b). The presence of

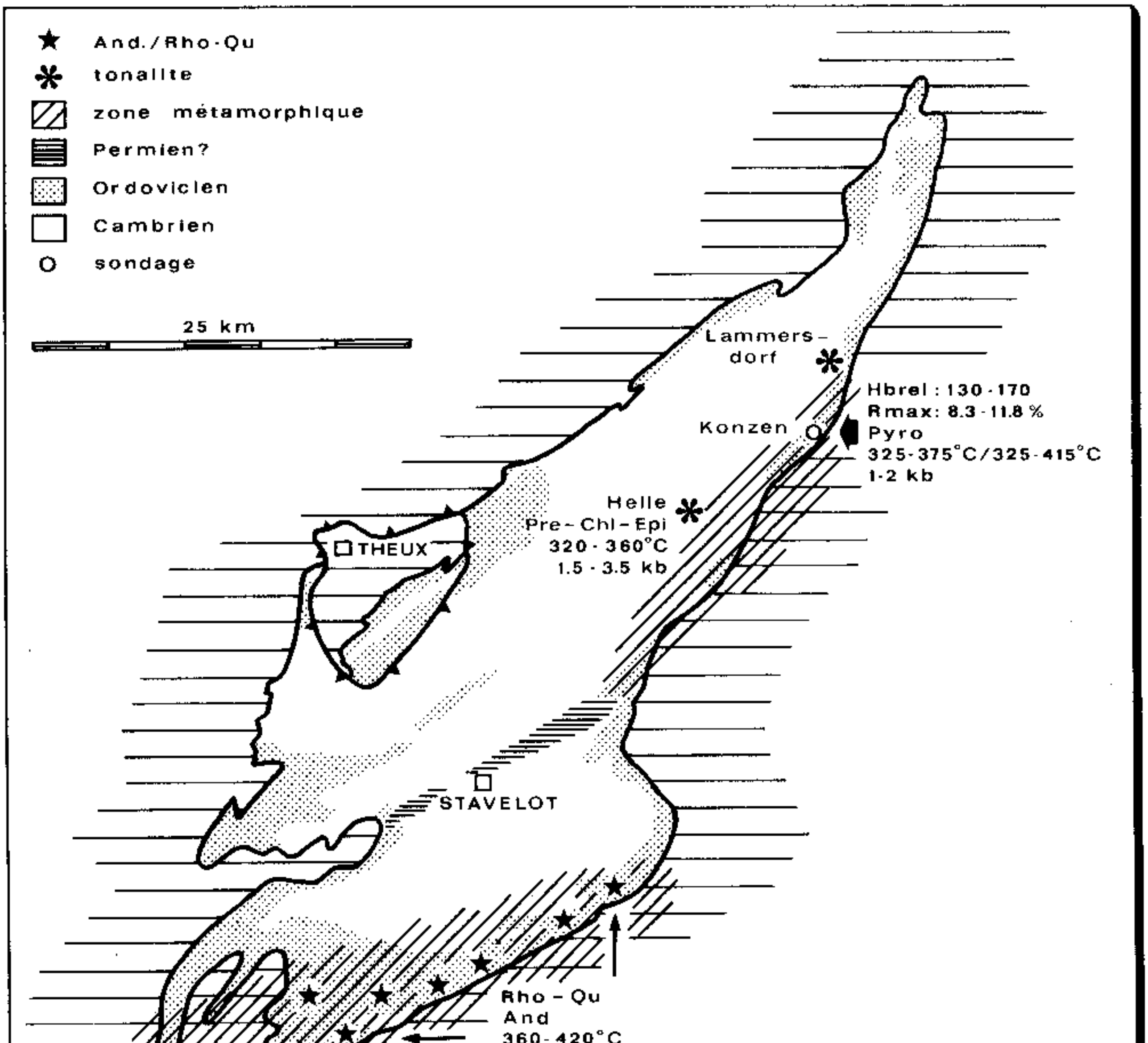


figure 8

MÉTAMORPHISME VARISQUE

sensitive minerals such as andalusite on the one hand and rhodochrosite and quartz on the other in the Salm deposits to the south of Stavelot suggests a maximum pressure of 2-3 kb and a formation temperature between 360 and 420°C (Kramm *et al.*, 1985b). The Upper Salm metapelites in the northeastern area (Konzen borehole) contain pyrophillite (Echle *et al.*, 1985), which indicates that metamorphism took place at 1-2 kb, and at temperatures of 325-375°C and 325-415°C, respectively (Kramm *et al.*, 1985b). Comparable temperature (320-360°C) and pressure (2.5 ± 1 kb) have been established for the prehnite-chlorite-epidote mineral assemblage of the Helle tonalite (Schreyer & Abraham, 1978).

Additional data on the Variscan metamorphism in the northeast part of the Stavelot Massif are provided by the study of illite crystallinity and coalification. High illite crystallinity ($Hb_{rel.}$ 100-120) in a narrow belt in the Cambro-Ordovician and Devonian rocks along the southeast flank of the massif near Konzen indicates the Variscan age of the low-grade metamorphism (temperature >350°C) in this area. Relatively low $Hb_{rel.}$ values in the Cambro-Ordovician rocks in the northwestern part of the massif may reflect a very low-grade Caledonian metamorphism that towards the east is overprinted by the Variscan metamorphism (Kramm *et al.*, 1985b). The discrepancy between the very high coal rank (R_{max} 8.3-11.8%) and the relatively low illite crystallinity ($Hb_{rel.}$ 130-170) in the Ordovician rocks of the Konzen borehole evidences the relatively short duration of the heat flow, because the improvement of illite crystallinity at a given temperature has a longer duration than the coalification of organic material (Kramm *et al.*, 1985b). These investigations should be extended to the Belgian part of the Stavelot Massif.

LATE CRETACEOUS TRANSGRESSION

(M.J.M.B., A.D., P.J.F., F.Gu., J.W.M.J., F.R.)

The residual Late Cretaceous deposits on the Hautes Fagnes massif evidence a step by step transgression of the area, which must have been a monadnock rising some 150 m above a pre-Late Cretaceous peneplain extending at least from the Stavelot area in the SE to the eastern Campine area in the NW (figs 9-10) prior to the Santonian to Maastrichtian transgression (Bless *et al.*, 1990). However, the southerly and easterly extension of the Campanian-Maastrichtian sea (perhaps flooding the Haute Ardenne and Eifel; cf. Demoulin, 1987a) remains a matter of debate.

«TONGRIAN» TRANSGRESSION

(M.J.M.B., A.D., C.D., P.J.F., F.Gu., J.W.M.J., A.P.)

The Cretaceous and Oligocene deposits in the Hautes Fagnes area are separated from those north of the line Liège-Aachen by an almost 20 km wide WSW-ENE directed depression with outcropping Paleozoic rocks (fig. 11). The rather deep valleys in this depression (e.g. those of the Inde, Vesdre and Amblève) came into existence during the Middle Oligocene when the Stavelot Massif and its northern foreland were slightly tilted to the NW, but the broad, asymmetric depression itself must have already been shaped by intensive erosion prior to the Oligocene. In the deepest part of this pre-Tongrian valley (about 200 m below the valley shoulders) not only the complete Cretaceous but also the upper portion of the underlying Paleozoic were removed. The northern valley slope consisted of eluvial flint covering the incompletely dissolved Cretaceous carbonates. Completely decalcified residual Cretaceous deposits were locally preserved on the crest of the Hautes Fagnes on the south flank of the valley. Through this valley, the sea invaded the region during the «Tongrian» (Early Oligocene; as explained by Gullentops, 1988, these strata are perhaps better placed in the Rupelian, since the main transgression took place during Rupelian time). The oldest «Tongrian» conglomerates and sands were deposited directly onto the Paleozoic basement (e.g. at Boncelles, and between 480 and 500 m altitude in the SW Hautes Fagnes at Minières/Tir Communal, Rondfahai and Trou Laurent; cf. Demoulin, 1987b, 1989).

Subsequently, the «Tongrian» transgression encroached upon the valley slopes and the northern plateau. On the lower part of the northern slope the so-called «Holset» sands and conglomerates were deposited on the eluvial flint. «Tongrian» sand with an occasional thin basal conglomerate rests on top of eluvial flint (e.g. at Halembaye; cf. Albers & Felder, 1981) or on the karstified top of the youngest Maastrichtian or Dano-Montian carbonates (e.g. at Enci and Thermae). More or less complete «Tongrian» sequences are preserved to the north of this pre-Tongrian valley (e.g. at Kleine Spouwen, Bunde, north of Valkenburg, and in the shafts of former coalmines in southern Limburg; cf. Albrecht & Valk, 1943; Felder *et al.*, 1984; Hinsberg *et al.*, 1973; Kuyl, 1980). In ascending order these consist of the Grimmertingen Sand, Neerrepen Sand, Henis Clay and Oude Biezen Sand, which all point to a very shallow shelf sea environment with transitory emersion (e.g. Neerrepen paleosol; cf. Gullentops, 1988).

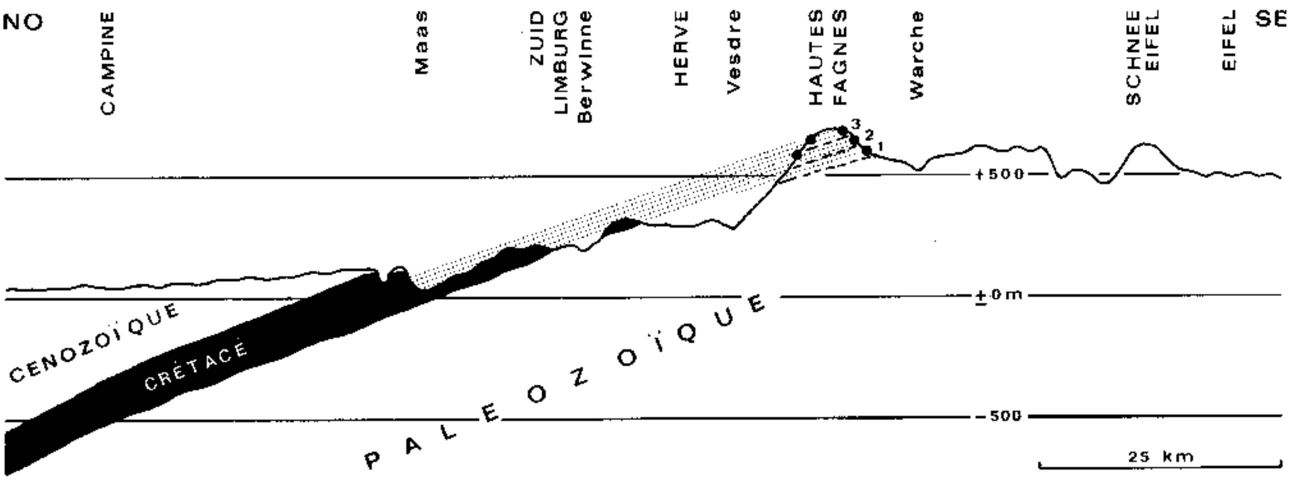


Figure 9.- NW-SE cross section showing step-like topography of Belgian Ardenne and their foreland, which developed independently after deposition of Late Cretaceous strata. Black dots in Hautes Fagnes relief indicate residual Late Cretaceous sediments. Figures 1 to 3 mark step by step flooding of Hautes Fagnes massif.

1. sequences starting with Early Campanian basal conglomerate; 2. Late Campanian basal conglomerate; 3. Late Maastrichtian basal conglomerate.

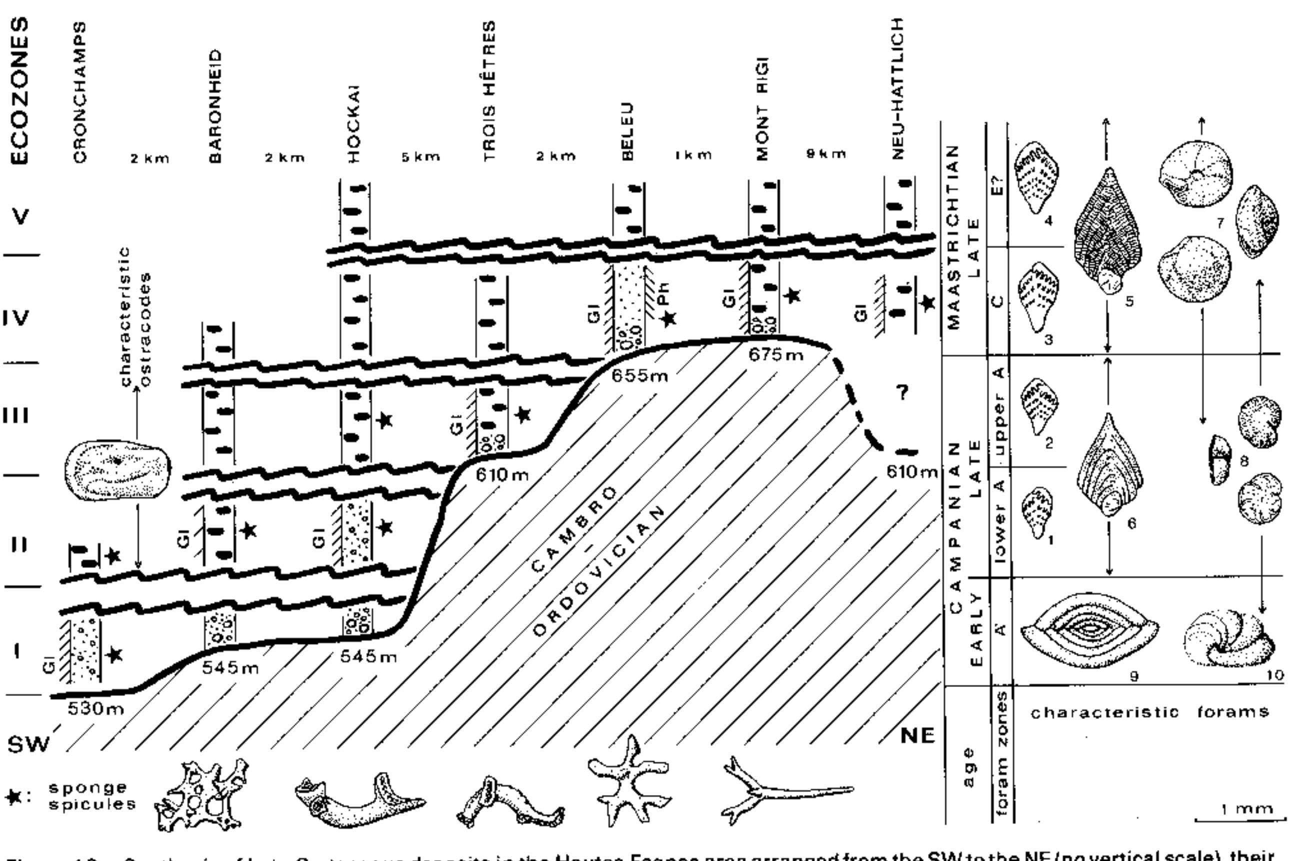


Figure 10.- Synthesis of Late Cretaceous deposits in the Hautes Fagnes area arranged from the SW to the NE (no vertical scale), their stratigraphic interpretation and some characteristic stratigraphic (forams, ostracodes) and paleoecological (sponges) elements. Note close correlation between transgressive trend of successively onlapping Cretaceous deposits and present-day height above sea level (indicated underneath each section) suggesting that actual morphology (at least from the SW to the NE) already existed during the Late Cretaceous.

Characteristic ostracodes : Late Campanian in the Hautes Fagnes is characterized by frequency and diversity of genus *Cytherelloidea*; the same is true for southern Limburg and Campine area. Presumably, this genus indicates a warming phase of the sea water (cf. Bless, 1989).

Characteristic forams : 1: *Bolivinoides* gr. *decoratus* (Jones) with predominantly three pustules on last chamber; 2: *B.* gr. *decoratus* with predominantly four pustules on last chamber; 4: *B.* gr. *australis* with predominantly six pustules on last chamber; 5: *Neoflabellina reticulata* (Reuss); 6: *N. leptodisca* (Wedekind); 7: *Eponides beisseli* Schijfsma; 8: *Stensioeina pommerana* Brotzen; 9: *Spirophthalmidium cretaceum* (Reuss); 10: *Ataxophragmium crassum* (d'Orbigny).

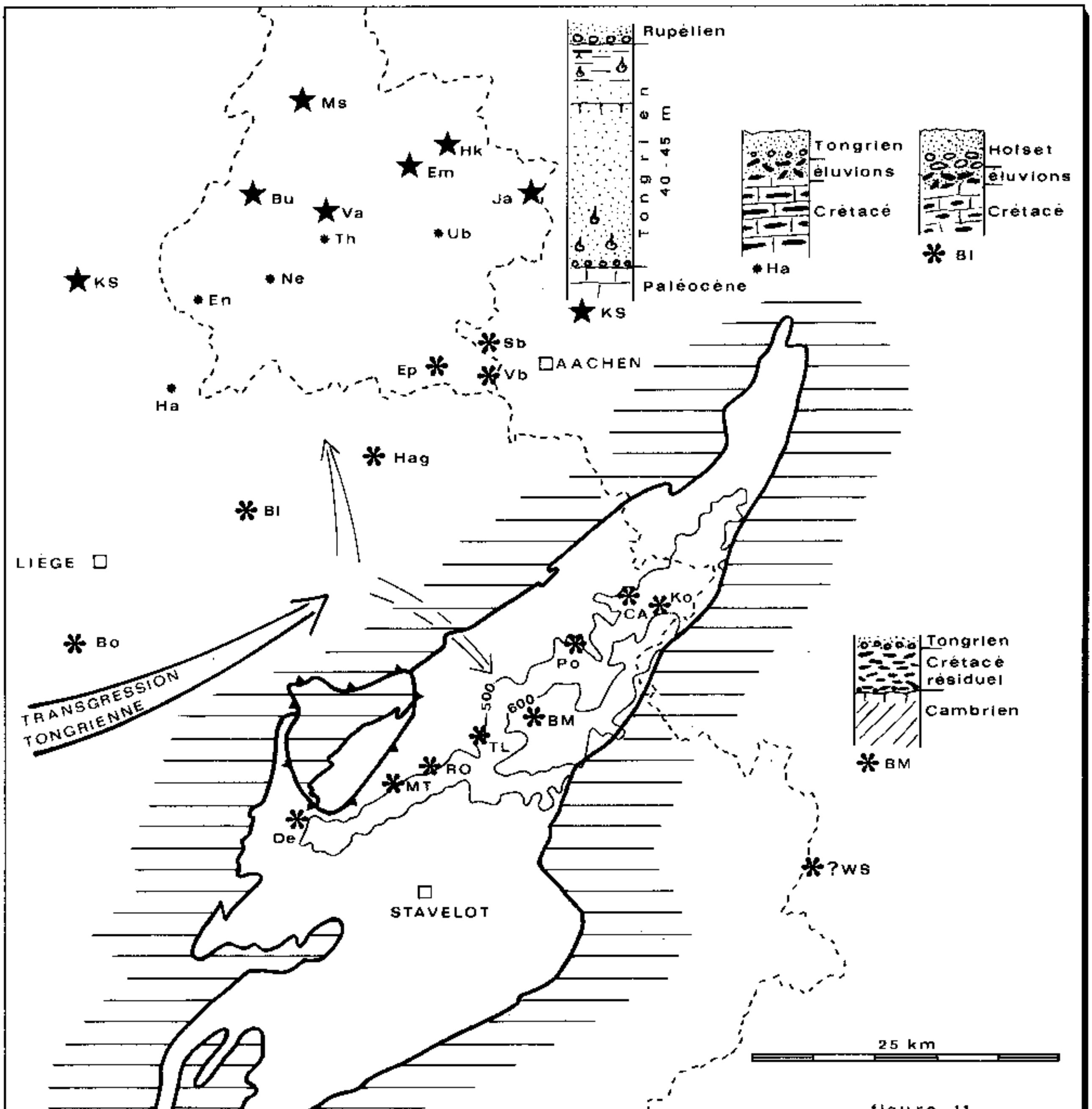
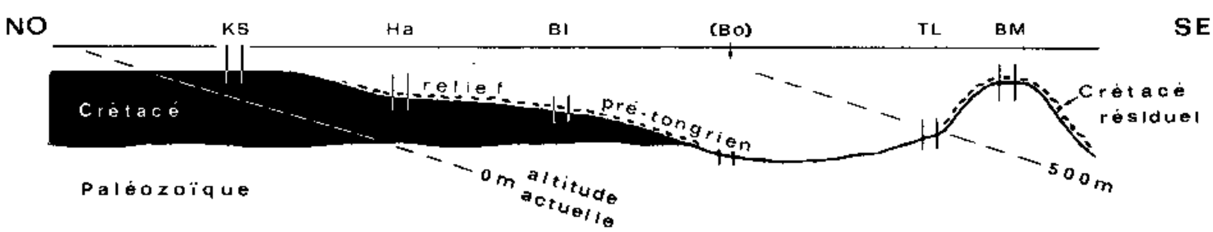


figure 11

TRANSGRESSION "TONGRIENNE"



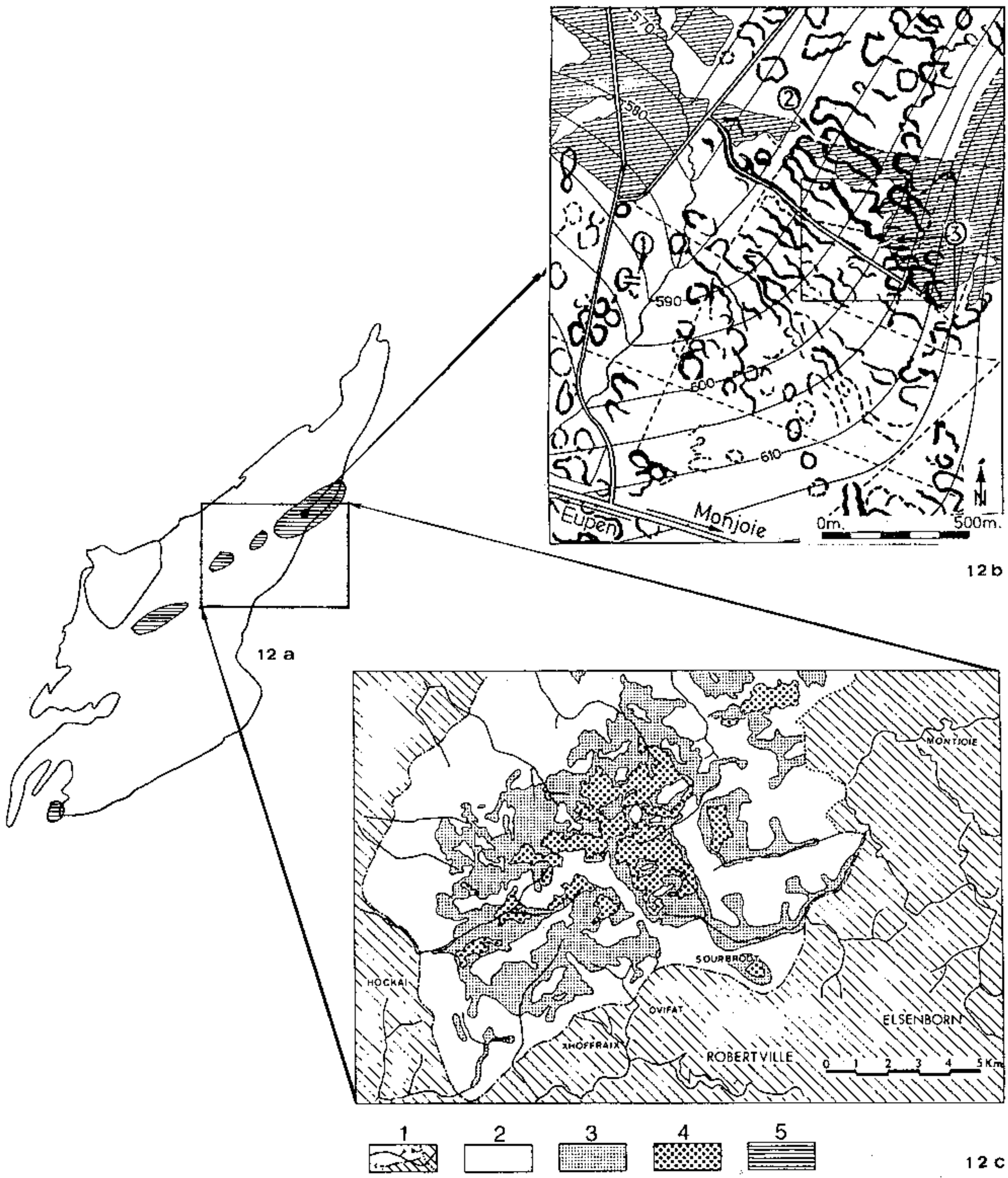


Figure 12.- Distribution pattern of palsas on the Stavelot Massif (a, b, after Pissart & Juvigné, 1980) and peat development in the Hautes Fagnes ares (c, after Dalimier *et al.*, 1985).
1a: brook; 1b: Belgian-German frontier; 1c: limit of the Hautes Fagnes area; 2: peat absent;
3: thin peat layer (0-80 cm); 4: thick peat layer (more than 80 cm); 5: palsas fields.

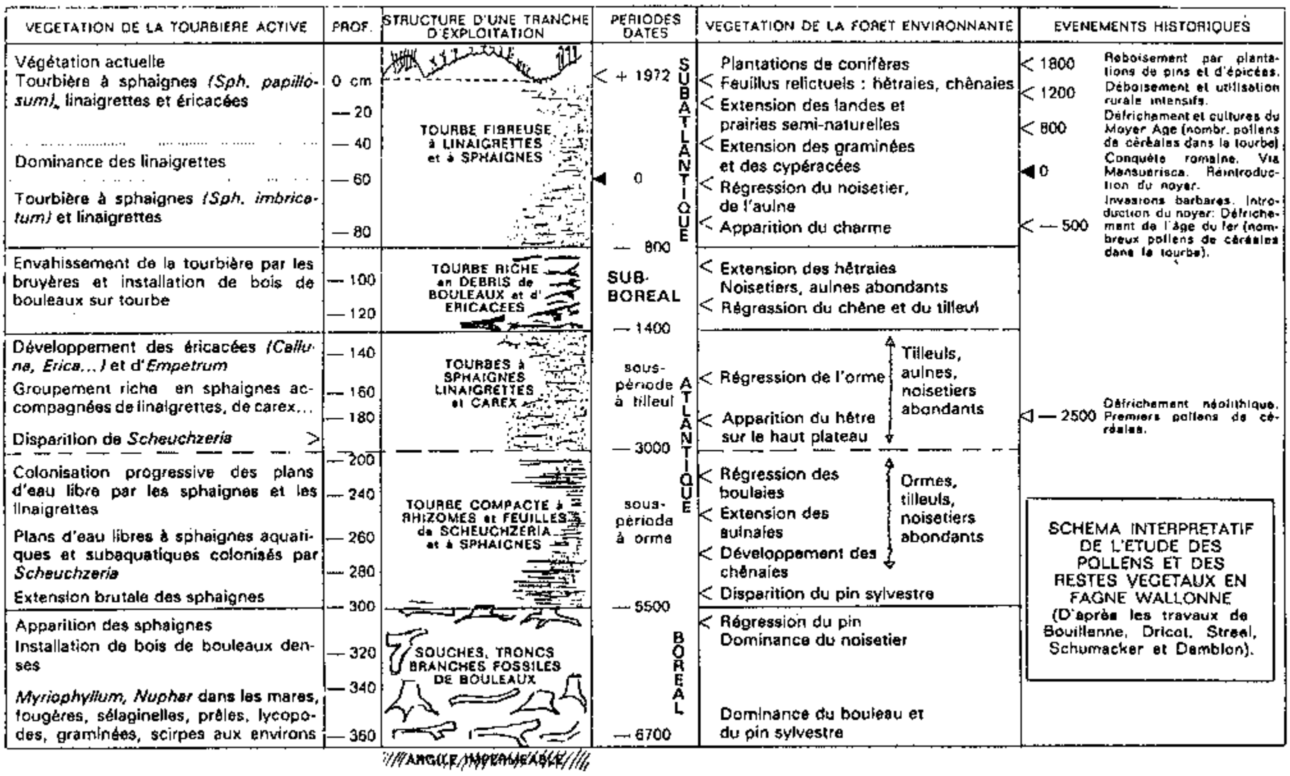


Figure 13.- Influence of climatic changes on the regional peat forming environment of the Hautes Fagnes during the Holocene (after Schumacker & Noirfalise, 1979).

Thin «Tongrian» deposits also occur on top of the residual Cretaceous at Baraque Michel on the summit of the Hautes Fagnes (Macar, 1954; Demoulin, 1989). The postulated southerly extension of the «Tongrian» sea into the Weisser Stein area (Demoulin, 1989) is in need of additional evidence.

QUATERNARY PEAT FORMATION

(E.J., A.P., R.S., M.S.)

Peat formation on the Hautes Fagnes has a long and very complex history. For example in the Brackvenn area (northeast part of the Hautes Fagnes), guttya and peat formation already took place during the Bölling and Allerød, but was interrupted during the Younger Dryas when hundreds of periglacial mounds «grew» on the plateau. Traces of these mounds are easily recognized in the field. They have been named «viviers» (Bouillenne *et al.*, 1937), pingos (Pissart, 1956), palsas (Pissart & Juvigné, 1980), and more recently mineral palsas (Pissart, 1983), respectively.

Mineral palsas are mounds which grow because of the development of ice lenses in the soil (segregated ice) and develop in periglacial environments where the mean annual temperature is slightly below 0°C (Dionne, 1978). In eastern Belgium, remnants of mineral palsas occur on plateaus above 550 m (fig. 12). They consist of circular to elongated peat bogs surrounded by rims of loose mineral material (mainly clayey silt including debris from local Paleozoic rocks). 14C dating, pollen diagrams and tephrostratigraphic data indicate that the palsas formed during the Younger Dryas between 10.700 and 10.000BP (Bastin, 1985; Bastin *et al.*, 1974; Gewelt, 1983; Mullenders & Gullentops, 1969; Pissart, 1983, 1985; Pissart & Juvigné, 1980).

During the Boreal, peat forming vegetation started to cover almost the entire Hautes Fagnes area above the 600 m level, filling depressions with up to 8 m thick Holocene peat. Palynological investigations (Persch, 1950; Streel, 1959; Damblon, 1978; Schumacker & Noirfalise, 1979) have revealed the main changes in climate and regional environment which affected the peat formation in this region (fig. 13). These studies need a further refinement. They are incorporated into a Global Change Programme. This would enable us to obtain a better understanding of the impact of man on these extremely vulnerable environments.

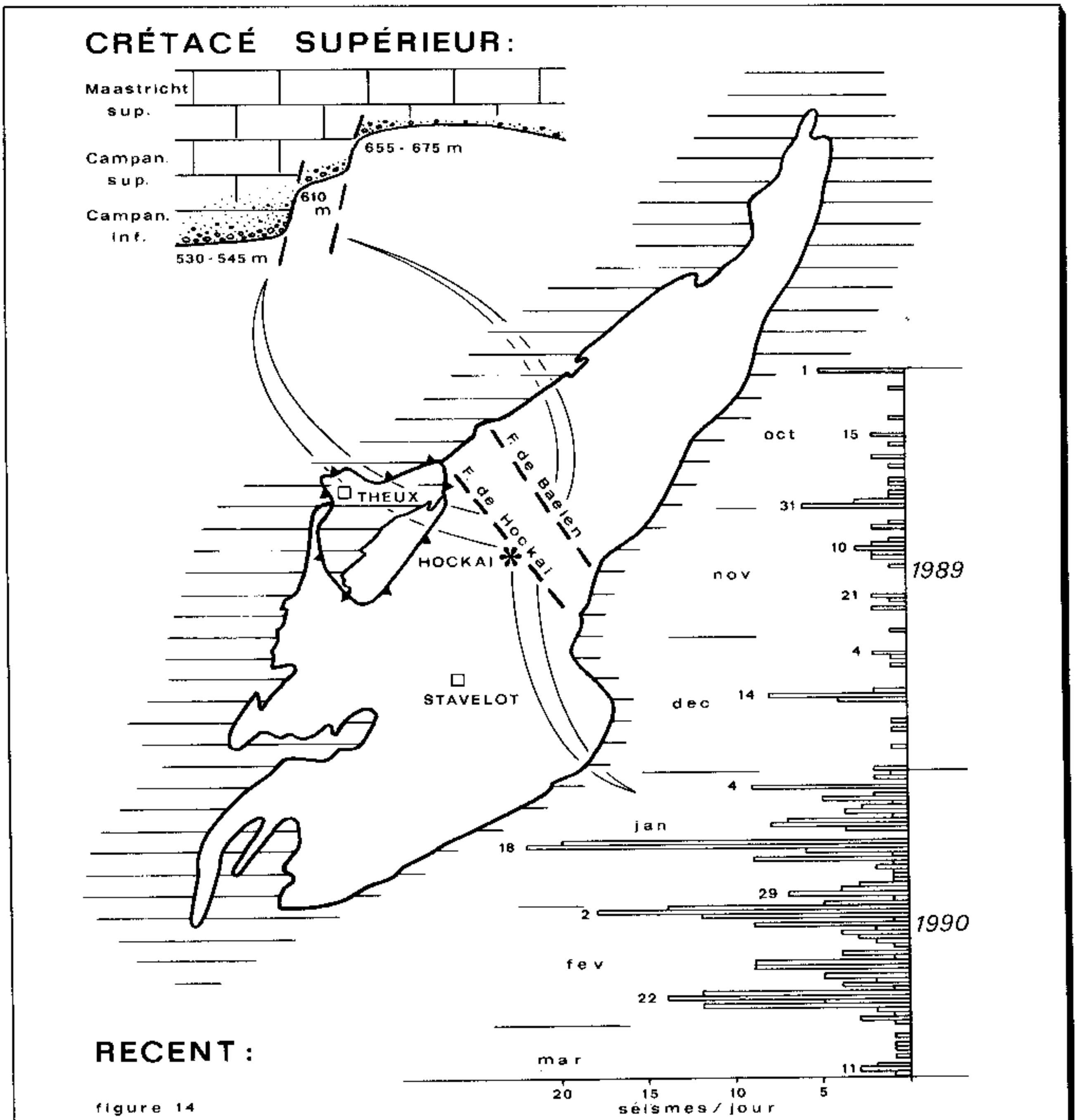


figure 14
FAILLES ACTIVES

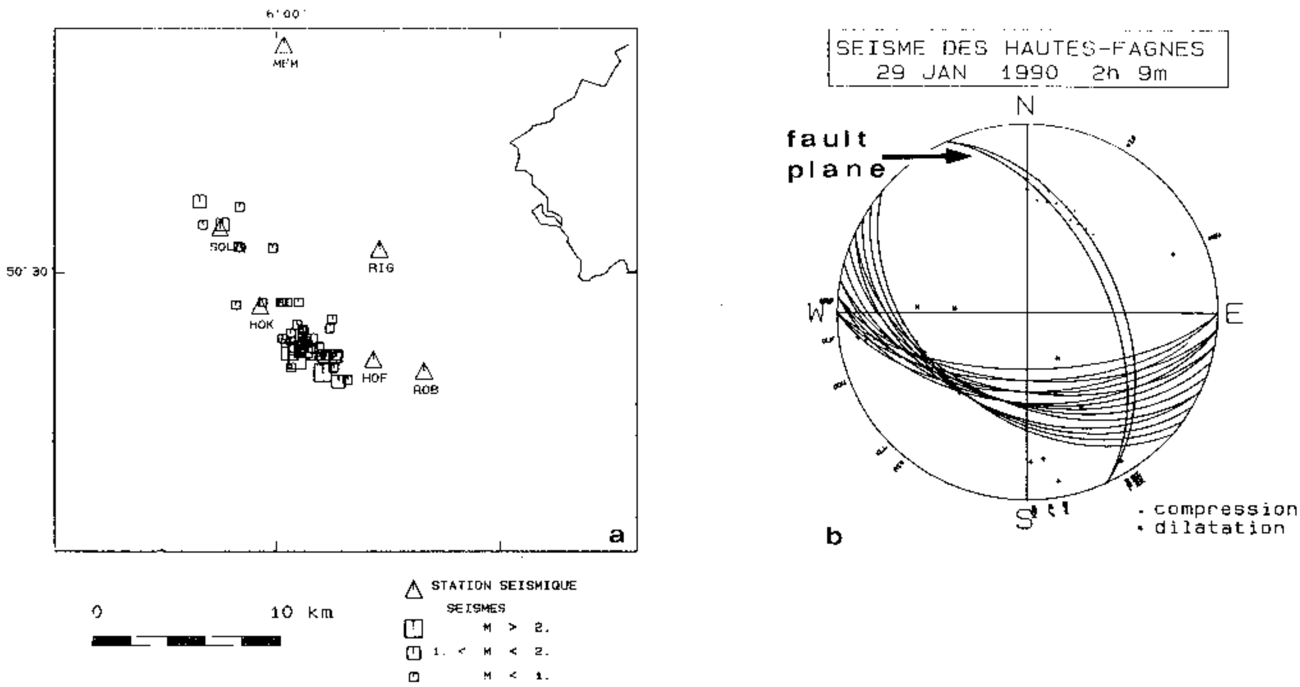


Figure 15.- Hypocentral repartition of earthquakes along Hockai fault zone and fault plane solutions of the Hautes Fagnes earthquake of January 29, 1990 (Camelbeeck, unpublished).
MEM: Membach; SOL: Solwaster; HOK: Hockai; RIG: Mont Rigi; HOF: Xhoffrai; ROB: Robertville.

QUATERNARY TECTONIC ACTIVITY

(M.J.M.B., J.B., T.C., L.D., A.D., F.Gu., L.H.)

Quaternary tectonic movements along NW-SE directed fault zones in the Stavelot Massif have been established by geomorphological investigations (Demoulin, 1988; Gullentops, 1987). At least some of these seem to have been active already during or prior to the Late Cretaceous (fig. 14; Bless *et al.*, 1990). Field identification of such transverse faults with limited throws is not easy due to the lack of lithologic contrasts. However they are clearly marked in the sedimentary cover on both sides of the Stavelot Massif (see tectonic map of Knapp, 1978). For example, the Hockai fault can be easily connected with the Verviers Fault (Fourmarier & Kolatchevsky, 1933).

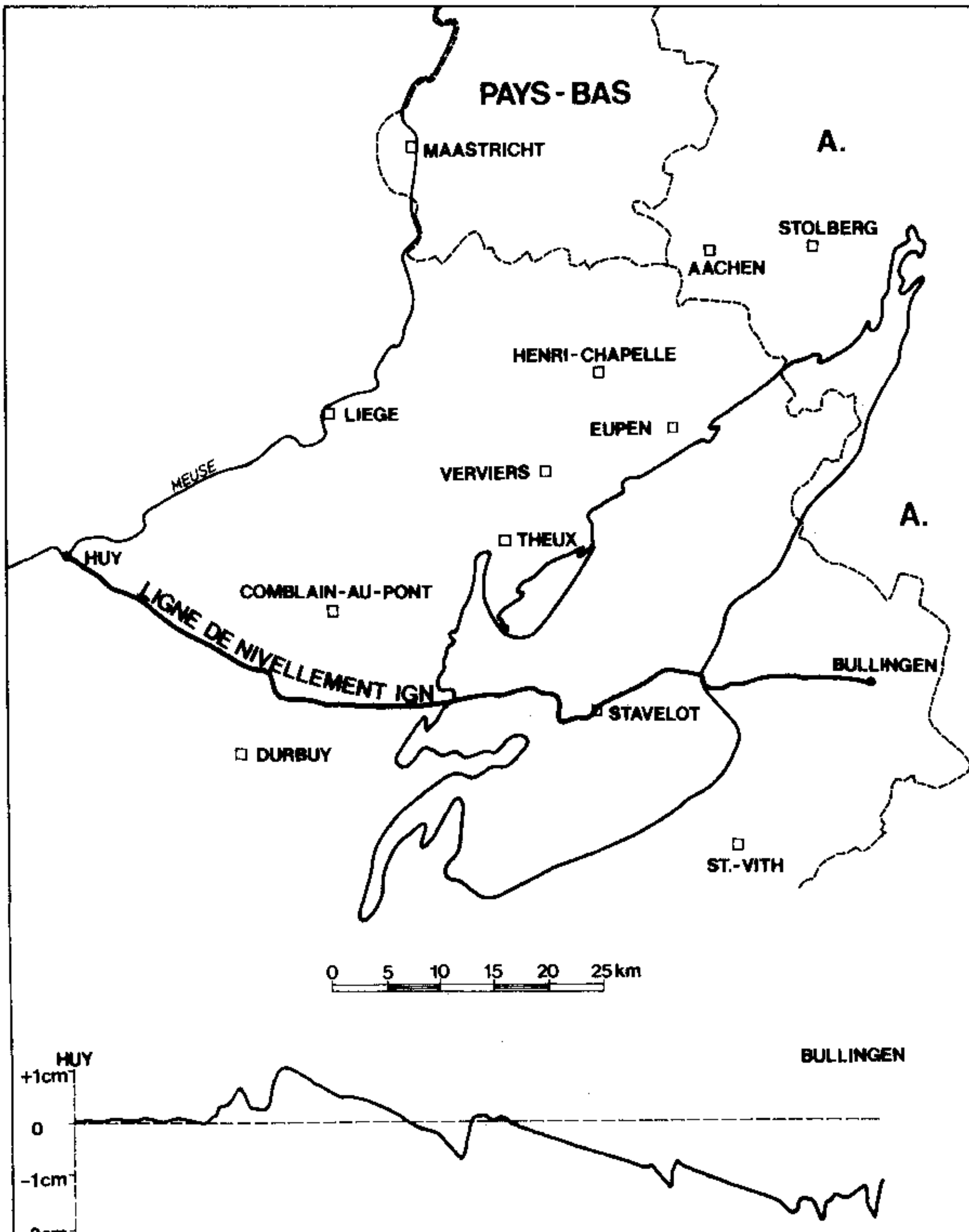
The Hockai fault zone (which may have experienced inverse movements in the past!) is still active as was illustrated by the seismic activity from October 1989 to April 1990 (fig. 14) and by the Malmédy earthquake ($M_l = 2.5$) of May 12, 1985 (Camelbeeck, 1990). The hypocentral repartition and the fault plane solutions (fig. 15) indicate activity along a normal NW-SE fault with a 60° dip to the NE, parallel to the faults in the Lower Rhine Embayment.

CONTEMPORARY CRUSTAL MOVEMENTS

(A.P., A.D.)

Comparison of precise levellings made in 1892 and in 1948 shows that the Hautes Fagnes plateau had been uplifted between these two dates. The amount of the uplift was 103 mm near the upper Amblève region, which corresponds with a very high mean speed of 2 mm of uplift per annum. From 1948 to 1980 the same locations settled down by more than 30 mm. Such important movements are typical of very unstable tectonic places.

The levelling line between Huy and Bullingen shows several blocks systematically tilted to the east (fig. 16). The limits of these blocks correspond with lines detected by remote sensing, geomorphology or seismic profiles.



Variations d'altitude par rapport au repère d'Uccle
de la ligne de nivellement de premier ordre Huy-Bullingen entre 1948 et 1980.

figure 16

RESUME

L'histoire géologique connue du Massif de Stavelot commence au Cambrien inférieur avec les dépôts de Deville. Elle est marquée par des lacunes importantes qui reflètent l'influence des grands mouvements tectoniques calédonien, varisque et alpin (fig. 17). Aujourd'hui cette histoire continue. Cela est démontré par le dépôt de tourbe, par les activités séismiques actuelles et par des variations d'altitude importantes pendant les dernières décades.

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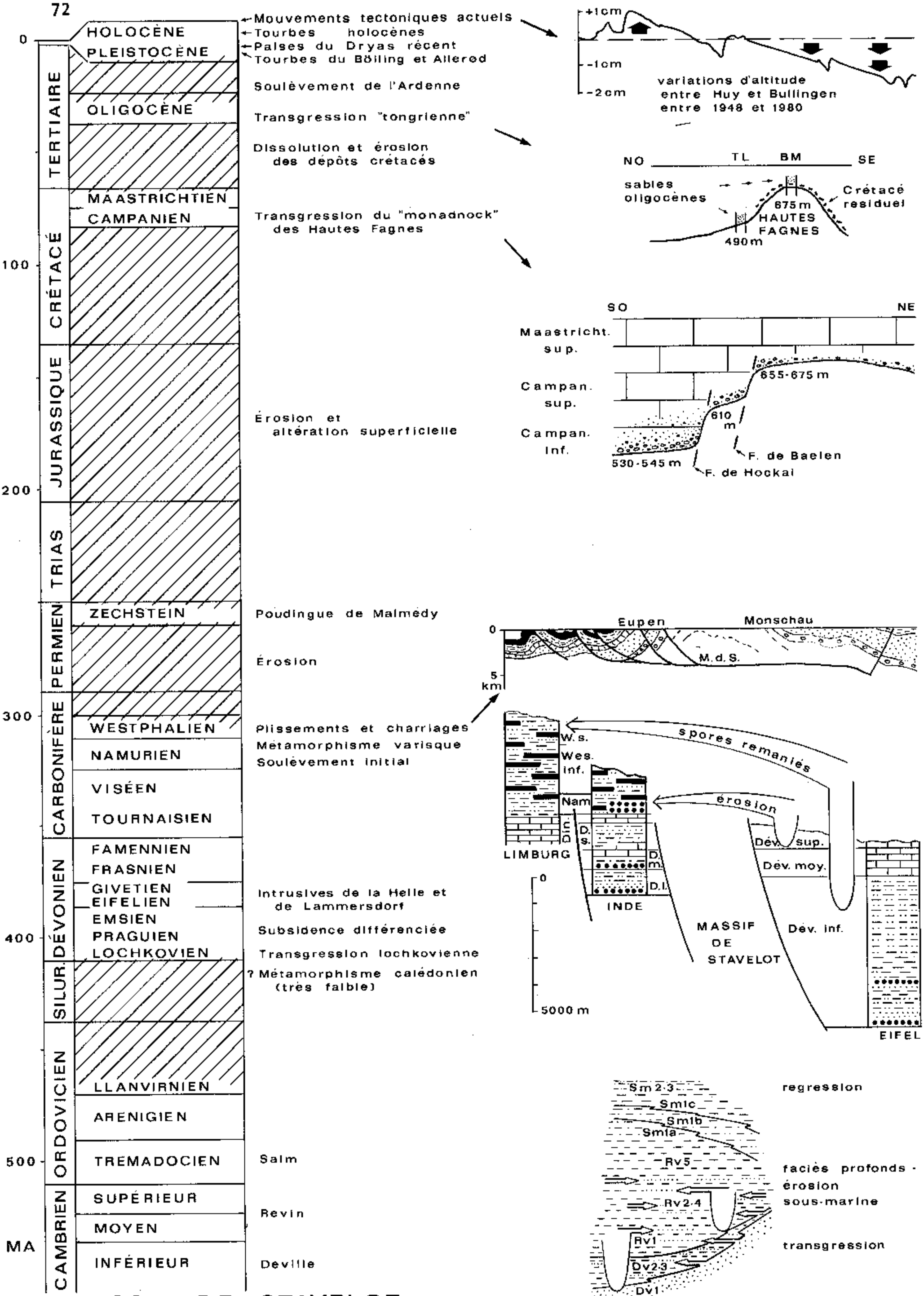

LE MASSIF DE STAVELOT

figure 17

KRAMM, U., BUHL, D. & CHERNYSHEV, I.V., 1985a. Caledonian or Variscan metamorphism in the Venn-Stavelot Massif, Ardennes? Arguments from a K-Ar and Rb-Sr study. *N. Jb. Geol. Paläont. Abh.*, 171: 339-349.

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