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Preliminary note on the organic facies, thermal maturity and dinoflagellate cysts of the Upper Maastrichtian Wang Formation in the northern subalpine massifs (Western Alps, France)

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Key-words: Dinoflagellate cysts, French Alps, Helvetic domain, Late Maastrichtian, Organic matter, Palaeoenvironment, Palynofacies, Prealpine nappes, Rock-Eval pyrolysis, Subalpine massifs, Thermal maturity, Ultrahelvetic nappes, Wang Formation.

ABSTRACT

Organic matter in the Upper Cretaceous Wang Formation was investigated in the northern subalpine massifs (Bornes, Bauges and Chartreuse) using Rock-Eval pyrolysis and palynological preparations. Results are threefold:

– Organic content is low (0–0.9% TOC) and consists mainly of type III organic matter, with a variable amount of dinoflagellate cysts and cutinite. This palynofacies indicates open marine, well-oxygenated depositional conditions.

– Thermal maturity tentatively derived from Rock-Eval T_{max} and dinoflagellate thermal alteration increases eastwards. This gradient may be interpreted, at least partly, as the result of burial under the now-eroded Ultrahelvetic and Prealpine nappes.

– In one of the thermally immature samples, a rich dinoflagellate cyst assemblage indicates a Late Maastrichtian age.

RÉSUMÉ

La matière organique de la Formation de Wang (âge Crétacé supérieur) a été étudiée dans les massifs subalpins septentrionaux (Bornes, Bauges, Chartreuse) à l'aide de la pyrolyse Rock-Eval et de préparations palynologiques. Les résultats suivants ont été obtenus:

– Le contenu organique est faible (0–0.9% COT) et consiste surtout en matière organique de type III, avec une proportion variable de kystes de dinoflagellés et de cutinite. Ce palynofaciès est caractéristique de conditions de dépôt de mer ouverte.

– Un essai d'évaluation de la maturité thermique à l'aide du Rock-Eval T_{max} et du degré d'altération thermique des dinoflagellés montre une augmentation de la maturité d'ouest en est. Ce gradient peut être interprété, au moins en partie, comme le résultat d'un enfouissement sous le front des nappes ultrahelvétiques et préalpines.

– L'un des échantillons thermiquement immature a livré un riche assemblage de kystes de dinoflagellés qui indiquent un âge Maestrichtien supérieur.

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ZUSAMMENFASSUNG

Das organische Material der oberkretazischen Wang-Formation wurde in den nördlichen subalpinen Massiven (Bornes, Bauges und Chartreuse) mittels Rock-Eval Pyrolyse und palynologischen Präparaten untersucht. Drei Resultate können festgehalten werden:

- Der organische Gehalt ist gering (0–0.9% TOC) und besteht vor allem aus organischem Material vom Typ III. Dinoflagellaten-Zysten und Cutinit treten in unterschiedlichen Mengen auf. Diese Palynofazies weist auf offenmarine, gut durchlüftete sedimentäre Bedingungen hin.
- Die thermische Maturität, vorläufig nachgewiesen durch Rock-Eval T_{max} und die thermische Alteration der Dinoflagellaten, nimmt von Westen nach Osten zu. Dieser Gradient kann, zumindest teilweise, als Resultat der Versenkung unter den jetzt erodierten ultrahelvetischen und präalpinen Decken interpretiert werden.
- In einer der thermischen immaturen Proben wurde eine reiche Vergesellschaftung von Dinoflagellaten-Zysten gefunden, welche dem Obermaastricht zugeordnet werden kann.

Introduction

The Upper Cretaceous Wang Formation was originally known in the northern subalpine massifs as "limestones and fetid shales" (Douxami 1881) or as "Jereminella limestones and black shales" (Moret 1934). The latter author already correlated these

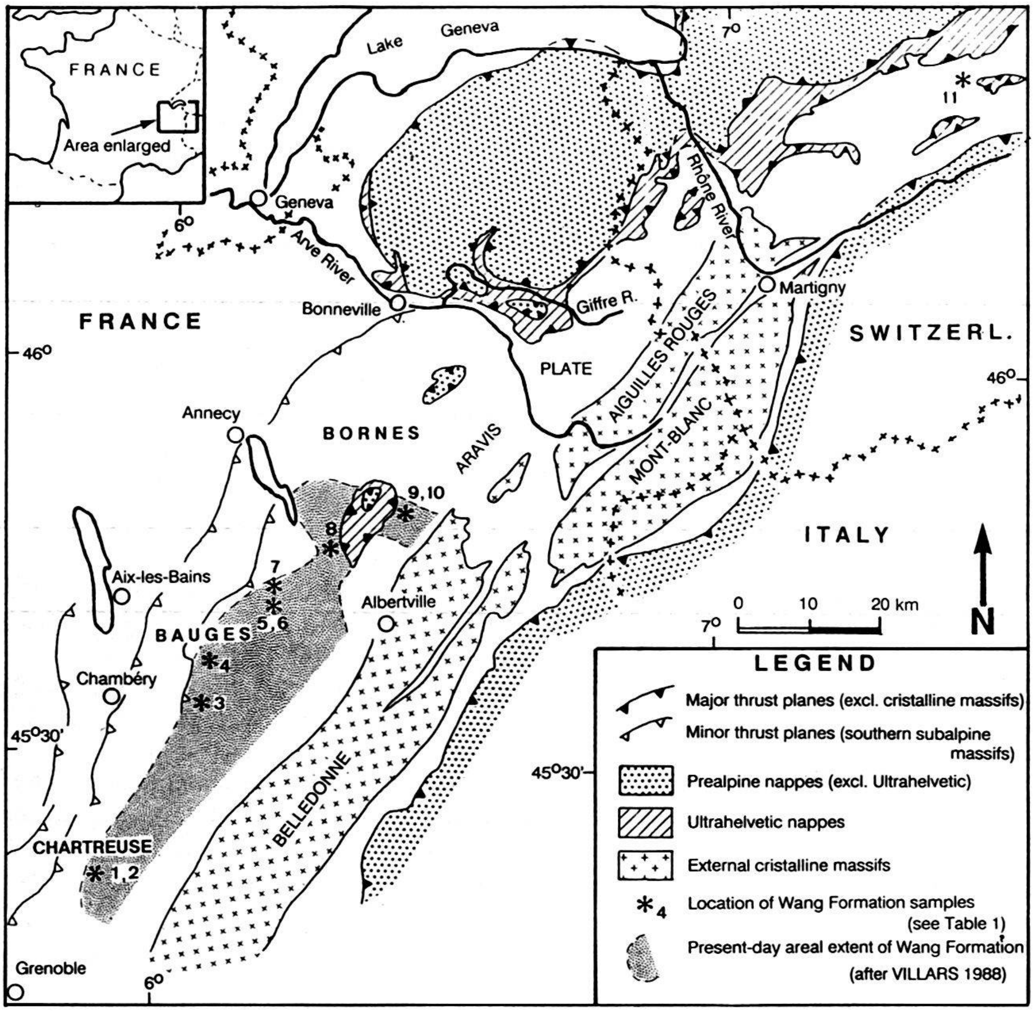


Fig. 1. Location map and tectonic framework.

deposits with the "Wang Layers" of the Helvetic and Ultrahelvetic nappes, but it is only in 1980 that Stacher formally defined the "Wang Formation" in the Helvetic nappes.

More recently, the Wang Formation has been studied by Villars (1988) in the northern subalpine massifs (Bornes, Bauges and Chartreuse, Fig. 1): it outcrops there as dm-bedded limestones, locally interbedded with marls. These rocks are finely laminated, locally bioturbated and characterized by their fetid smell when freshly broken. They are interpreted as deposited on an outer hemipelagic shelf with high sedimentation rates and are dated as Late Maastrichtian by planktonic foraminifera and nannofossils. The Wang Formation unconformably overlies Senonian marls and limestones and is truncated by the large-scale Early Tertiary erosion, which explains its present-day limited geographical extension (Villars 1988 and Figs. 1 and 2). In the Ultrahelvetic units between the Arve and Giffre Rivers to the northeast of the study area (Fig. 1), Kindler (1987, 1988) has studied "sublithographic" limestones dated as Lower-Middle Paleocene by planktonic foraminifera and compared with the Wang Formation. This structurally complex area has not been studied in this paper.

The purpose of the present study was to make a preliminary investigation of organic matter (OM) in the Wang Formation of the northern subalpine massifs. Ten samples (Fig. 1 and Table 1) were selected from different sections studied by Villars (1988). For the sake of comparison, a sample located in the Wang Formation of the Wildhorn Helvetic nappe was also analysed. Four Lower Tertiary samples studied by Gorin et al. (1989) in the Bornes Massif were also used as extra data points for thermal maturity. OM in Wang Formation samples was analysed by a geochemical method (Rock-Eval pyrolysis) and by microscopic observations of palynological slides. The results should be considered as preliminary because of the small number of samples analysed.

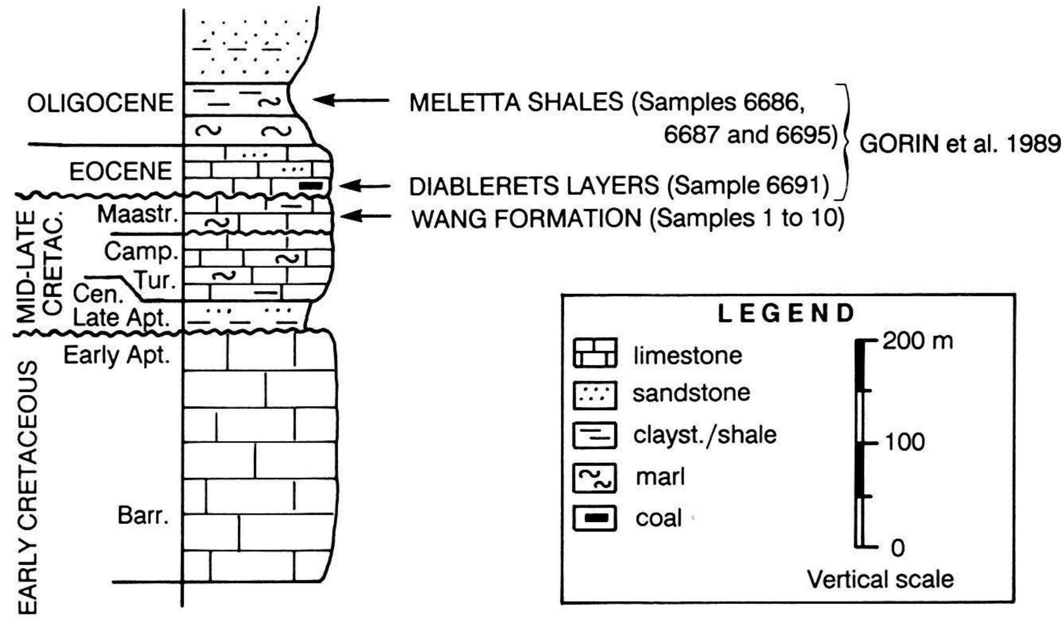


Fig. 2. Simplified stratigraphy of the Bornes-Bauges subalpine massifs (modified after Charollais 1988) and position of the Wang Formation and Lower Tertiary samples used in this study.

Sample nb.	Sample ref.	Section (after VILLARS 1988)	Lambert coordinates	Lithology	ROCK-EVAL RESULTS TOC(1) (% weight)	HI (mgHC/gTOC)	OI ($mgCO_2$/gTOC)	T_{max} (°C)
WANG FORMATION: NORTHERN SUBALPINE MASSIFS, FRANCE								
1	FV 1440	Col Bellefond	877.70/2044.95/1902 m	grey calc. marl	n/a	n/a	n/a	n/a
2	FV 1441	Col Bellefond	877.70/2044.95/1902 m	grey sil. lst.	< 0.10			
3	FV 1322	Les Fontanettes	890.80/2069.15/1230 m	brown recrist. lst	ca. 0			
4	FV 1227	Aillons le Jeune	892.40/2075.50/910 m	grey-brown sil. lst.	0.37	(140)	(200)	(429)
5	FV 1337	Pleuvens	900.40/2083.90/1950 m	grey arg. lst.	0.22	(172)	(240)	(430)
6	FV 1340	Pleuvens	900.40/2083.90/1950 m	grey arg. lst.	0.36	(230)	(166)	(431)
7	FV 1000	Dent des Portes	899.80/2085.10/1795 m	grey arg. lst.	0.63	185	104	430
8	FV 655	Roc de Viuz	907.25/2092.35/820 m	grey lst.	0.89	201	100	437
9	FV 780	Mont Charvin	917.45/2097.50/2150 m	black calc. clay	0.66	19	78	461
10	FV 785	Mont Charvin	917.45/2097.50/2150 m	grey-black marl	0.89	12	88	474
WANG FORMATION: SWISS ALPS								
11(2)	FV 1448	Luton (Wildhorn nappe)	7°27'E/46°21'30''N (geogr. coordin.)	grey arg. lst	0.18			
LOWER TERTIARY SAMPLES: BORNES SUBALPINE MASSIF, FRANCE (see GORIN et al. 1989)								
6686		Dessy	912.50/2124.84	grey-blue marl	1.27	323	0	442
6687		Dessy	912.50/2124.84	grey-blue marl	1.18	254	6	440
6691		Cenise	916.20/2119.40	black coaly lst.	7.42	169	0	460
6695		Nanoir	899.91/2109.60	grey-blue marl	1.32	400	15	428

n/a = not available, arg. = argillaceous, calc. = calcareous, lst. = limestone, sil. = siliceous

(1) Except for sample 4, samples with TOC values <0.25% give totally unreliable HI, OI and T_{max} values. HI, OI and T_{max} values of samples 4, 5 and 6 should be used with caution because of their TOC <0.5% (possible effects of mineral matrix, see Espitalie et al. 1985).

(2) Sample outside of the study area, mentioned only for the comparison of thermal maturity.

Table 1: Location, lithology and Rock-Eval results of samples studied in the Wang Formation. Lower Tertiary samples from the Bornes Massif (GORIN et al. 1989) are used in this paper as extra data points for thermal maturity (see Figs. 5 and 6).

Organic facies

Methods

Rock-Eval pyrolysis of total rock (Espitalie et al. 1985–1986) provides quick determination of type and richness of OM, using parameters such as TOC (total organic carbon), HI (hydrogen index) and OI (oxygen index).

Typing of OM can be further refined by microscopic studies of palynological residues, where OM in rock samples has been concentrated by elimination of the mineral phase using hydrochloric and hydrofluoric acids. This organic residue is not oxidized for the study of palynofacies and thermal alteration of OM. Nature and amount of figured and amorphous OM is evaluated in order to define the palynofacies (or organic facies) of the rock, which provides valuable clues to the depositional environment and brings useful complements to standard sedimentological and micropalaeontological observations (Batten 1982, Habib 1983, Hart 1986, Bustin 1988 and Gorin et al. 1989).

Results (Figs. 3 and 4)

The Wang Formation has a low content in OM, as indicated by Rock-Eval TOC values varying between 0 and 0.89% (Table 1). When using Rock-Eval parameters (Fig. 3), one should keep in mind the possible effects of mineral matrix (Espitalie 1985): HI values may be underestimated and T_{max} values overestimated when rocks

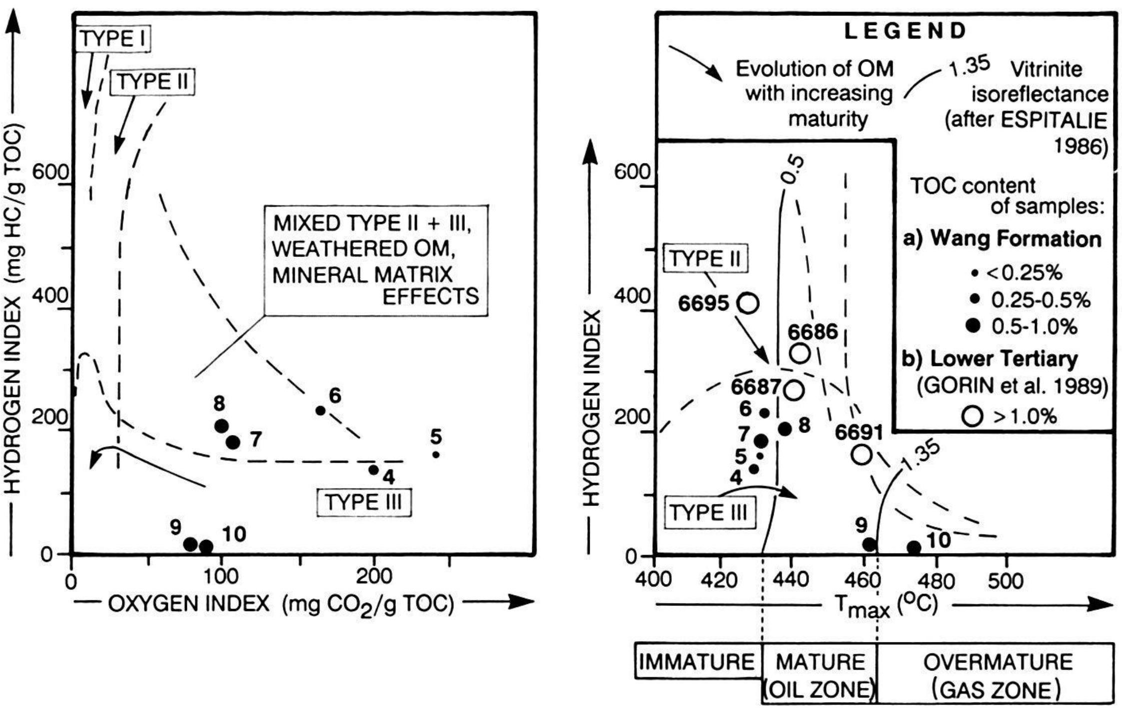


Fig. 3. Rock-Eval pyrolysis results: classification of organic matter in a HI-OI diagram and display of thermal maturity variations in a HI-T_{max} diagram. Mineral matrix effects may affect HI, OI and T_{max} values of samples 4 to 10. In particular, OI for samples 4, 5 and 6 may be too high because of their low TOC content (<0.5%).

are argillaceous, immature and poor in OM (TOC<1%) and OI values may be too high in organic-poor rocks (TOC<0.5%). The latter may be particularly true for samples 4, 5 and 6 which contain less than 0.5% TOC (Table 1). Recorded HI values rarely exceed 200 and all data points plot as type III OM (Tissot & Welte 1984) in a Van Krevelen diagram (Fig. 3), with possibly some admixture of type II OM. The high thermal maturity of samples 9 and 10 (see below) explains their low HI and OI. From these raw (i.e. uncorrected for mineral matrix effects) pyrolysis results, one would expect palynofacies of the Wang Formation to be dominated by humic fragments (inertinite and vitrinite) with some addition of liptinitic (i.e. hydrogen-rich) material, either of continental (cutinite, pollens, spores) or marine (phytoplankton) origin.

Fig. 4 and Plate 1 (Figs. 1 to 3) illustrate the palynofacies of the Wang Formation. Microscopic investigations confirm Rock-Eval predictions: the organic residue contains 50–80% of continental fragments, mainly hydrogen-poor vitrinite and inertinite, with locally some addition of hydrogen-rich cutinite. The marine influence is nevertheless pronounced, with only very few pollens and spores but numerous dinoflagellate cysts (8–45%). Cyst species are many (more than 30 in sample 1 studied in detail, see below) and evenly spread between proximate and chorate species. In absolute numbers there are on average twice as many chorate specimens than proximate ones. Foraminifera linings are also present in minor proportions.

Inertinite and vitrinite, derived from ligno-cellulosic compounds by oxidation and condensation are the most stable macerals in the organic residue (Hart 1986). On average their fragments have a smaller size and are less angular in the south (samples 1 to 4, Plate 1, Fig. 1) than in the north (samples 5 to 10, Plate 1, Figs. 2 and 3). This holds true in samples 8 to 10 despite thermal degradation (Plate 1, Fig. 3).

Because of the high thermal maturity, it is difficult to evaluate the palynofacies of sample 11 collected in the Wildhorn Helvetic nappe (Fig. 1). It contains only inertinite fragments and small blackened amorphous debris, most of which are probably remnants of palynomorphs, as indicated by ghosts of dinoflagellate cysts or pollens (Plate 1, Figs. 12 and 13).

In the thermally immature to marginally mature samples 1 to 7 (Plate 1, Figs. 1 and 2), all OM particles are well preserved and present little evidence of fungal or bacterial attack.

Discussion

The absence of a sapropelic amorphous groundmass (OM type I, Tissot & Welte 1984) and the abundance and variety of dinoflagellate cysts point to open marine, well-oxygenated palaeoenvironmental conditions, which are quite compatible with those of an "outer hemipelagic shelf" postulated by Villars (1988). Plankton is diluted by a terrestrial input made of inertinite, vitrinite and accessorily cutinite. Because of the limited number of samples it is premature to draw palaeogeographical conclusions from variations in size of these continental fragments.

The excellent preservation state of the organic particles in the immature to marginally mature samples can be easily explained by the high sedimentation rates calculated by Villars (1988): OM would have passed rapidly through the sediment-water interface and the upper layer of sediments, where OM degradation is the most intense

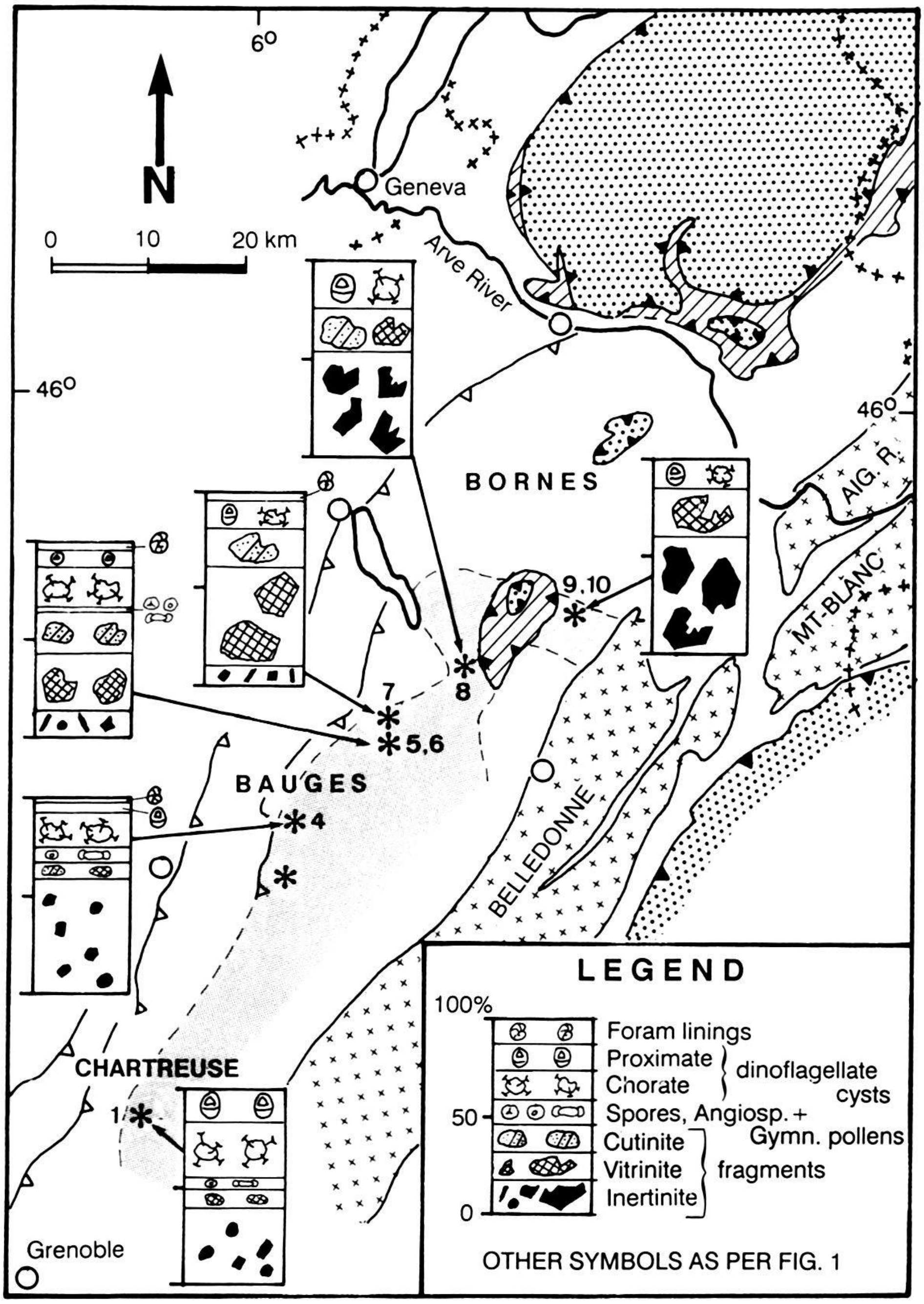


Fig. 4. Palynofacies distribution in the Wang Formation. Inertinite and vitrinite fragments in samples 5 to 10 in the north of the studied area are larger and more angular than those in samples 1 to 4 in the south.

(HART 1986). High sedimentation rates associated with a normal rate of organic production may also explain the low OM content of the Wang Formation.

Finally, the low OM content seems to indicate that the fetid smell of the Wang Formation is not linked to organic richness. Nevertheless, the presence of hydrogen sulfide in such a carbonate environment may result from the degradation of OM by sulfate-reducing bacteria (HART 1986).

Thermal maturity and possible implications for the front of the Ultrahelvetic and Prealpine nappes

Methods

The absence of coaly fragments in the Wang Formation prevents determination of thermal maturity by vitrinite reflectance. Alternative methods such as Rock-Eval pyrolysis and microscopic observations of OM may provide valuable, though less accurate, evaluations of OM maturation. The Rock-Eval T_{max} parameter, computed from temperature measurements using the kerogen cracking peak, provides an index of OM maturation (ESPITALIE 1986, ESPITALIE et al. 1985–1986, CRUMIERE et al. 1988). As mentioned above, T_{max} values may be overestimated in organic-poor rocks because of mineral matrix effects (samples 4, 5 and 6, Table 1). Rock-Eval analyses on more samples from the same location, or several runs on the same samples would be needed to obtain highly reliable results.

Microscopic observations of colour changes in unoxidized macerals is another way of estimating thermal maturity of OM and is referred to as thermal alteration index (=TAI, STAPLIN 1969). Although the TAI was originally defined on spores, the same change of colour can be observed on other liptinitic (i.e. hydrogen-rich) palynomorphs such as pollens and dinocysts (ROBERT 1985). Qualitative microscopic observation of the fluorescence of liptinitic unoxidized macerals (such as dinoflagellate cysts) in UV light may give another estimate of the thermal maturity of sediments (ROBERT 1985, GORIN et al. 1989). Because of the varying nature of dinocyst walls, observation of TAI or UV-fluorescence have be made on the same palynomorph species when comparing different samples.

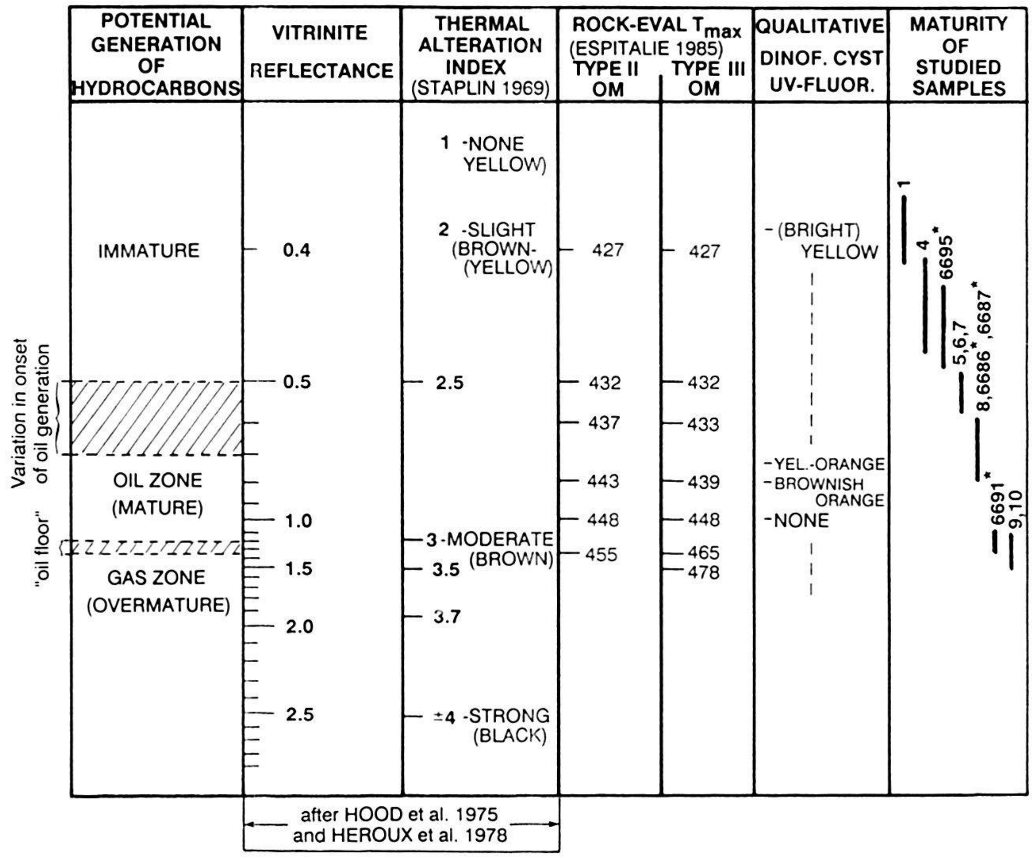
A tentative correlation between these various maturity parameters, the vitrinite reflectance (VR) scale and the different maturity zones of OM is shown on Fig. 5. We shall use the following threefold subdivision when referring to maturity of sediments (see also GORIN et al. 1989):

- immature VR<0.5%
- mature VR=0.5 to 1.35 (oil zone)
- overmature VR>1.35% (gas zone).

On Fig. 5, T_{max} values are correlated with VR on the basis of diagrams shown in ESPITALIE (1985, p. 765). The qualitative UV-fluorescence index for dinoflagellate cysts is derived from personal observations on unoxidized palynological slides prepared from rocks calibrated by vitrinite measurements. This qualitative index is comparable with that shown by ROBERT (1985, p. 176–177) for *Tasmanites* algae.

Results (Figs. 5 and 6)

Rock-Eval results are listed on Table 1 and variations in T_{max} values illustrated on a HI-T_{max} diagram (Fig. 3). T_{max} values are plotted on Fig. 6, together with TAI and UV-fluorescence of dinoflagellate cysts. Considering these three maturity parameters and their uncertainties, one can only define for each sample a tentative maturity range rather than a precise value (Fig. 5). Using the correlation shown on Fig. 5, maturity ranges can then be expressed in VR equivalent (Fig. 6). Uncertainties in VR equivalent values are highlighted on Fig. 6 by question marks.



*Lower Tertiary samples studied in GORIN et al. 1989.

Fig. 5. Scales of organic metamorphism and potential hydrocarbon generation (modified from GORIN et al. 1989). The maturity of Wang Formation and Lower Tertiary samples, derived from TAI, Rock-Eval T_{max} and UV-fluorescence, is plotted as a range and can be tentatively correlated with the vitrinite reflectance scale. Maturity of Lower Tertiary samples is also supported by biomarkers (GORIN et al. 1989).

The three methods used suggest an increase in thermal maturity from west to east. Samples 1 to 7 in the Chartreuse and Bauges Massifs are immature to marginally mature and samples in the Bornes Massif are mature (sample 8) to highly mature (samples 9 and 10). No T_{max} values are available for samples 1 and 2. Both contain dinoflagellate cysts which seem immature under the microscope (light yellow in natural light, see Plate 1, Fig. 4, and with a yellow UV-fluorescence). Samples 3 to 7 are more mature and lie close to the immature-mature boundary line as shown by T_{max} values and the stronger shade of yellow of dinoflagellate cysts in natural light (Plate 1, Figs. 5 to 8). All parameters show sample 8 to be mature: in natural light dinoflagellate cysts are clearly darkened (Plate 1, Fig. 9). Samples 9 and 10 have a higher maturity level, close to the mature-overmature boundary, as indicated by very high T_{max} values and by thermal degradation of dinoflagellate cysts (Plate 1, Figs. 10 and 11).

Extra maturity data in the same area have been obtained by GORIN et al. (1989) on four organic-rich (i.e. with a TOC content >1%) samples in the Bornes Massif, three from the Lower-Middle Oligocene Meletta Shales and one from the Upper Eocene Diablerets Layers. Maturity estimates of these samples are quite reliable because they are based not only on methods described above, but also on other geochemical ana-

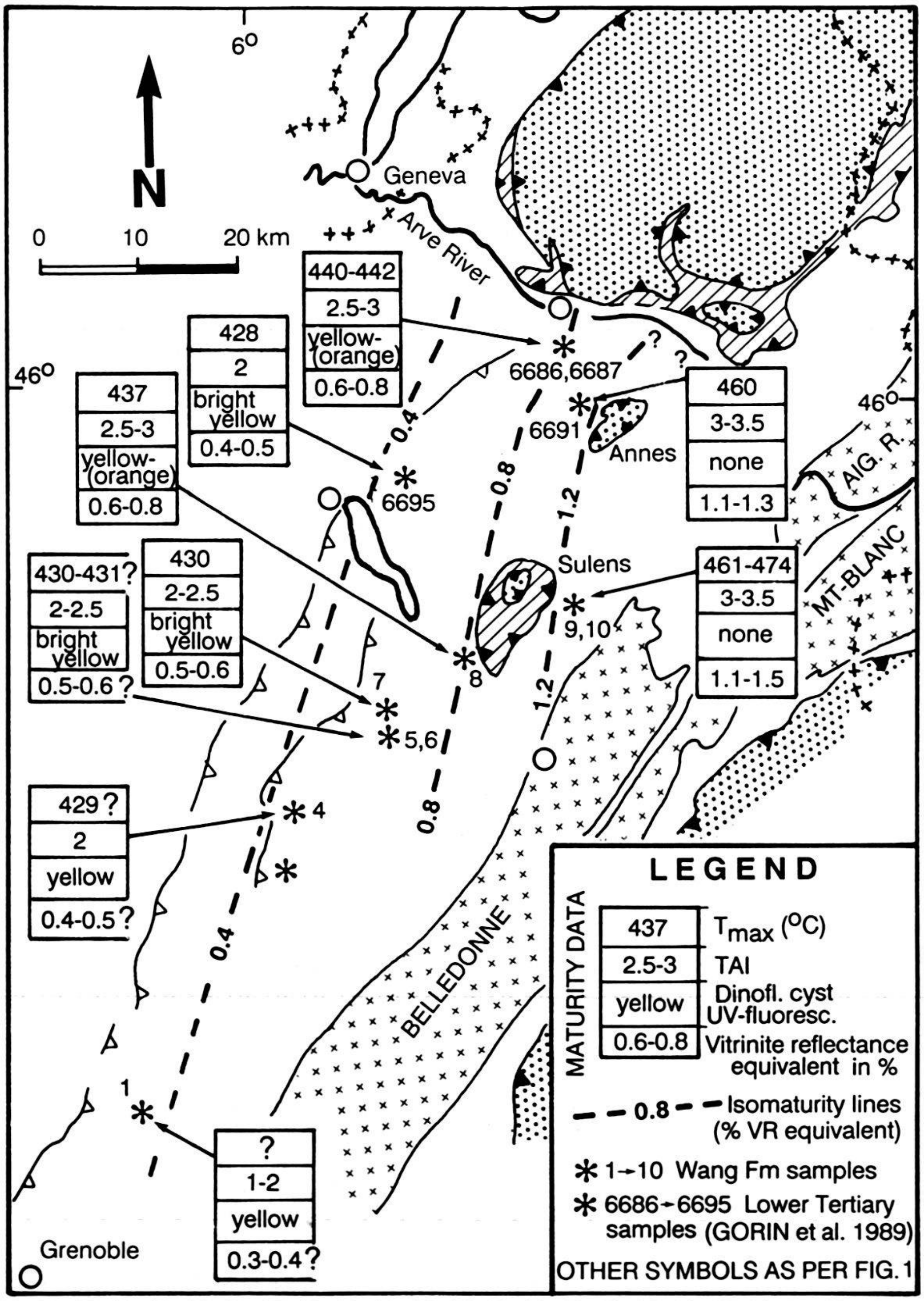


Fig. 6. Summary of maturity data of Wang Formation and Lower Tertiary samples. Tentative maturity map for Upper Cretaceous and Lower Tertiary sediments belonging to the Helvetic domain in the northern subalpine massifs. Question marks highlight uncertainties in VR equivalents for samples without T_{max} (sample 1) or with less reliable T_{max} because of low TOC content (samples 4, 5 and 6).

lyses (biomarkers). Because these samples are stratigraphically not too distant from the Late Maastrichtian Wang Formation (Fig. 2), their data can be added to those obtained for the Wang Formation and plotted in the same way on Figs. 5 and 6. Consequently, keeping in mind the possible uncertainties attached to some samples and the limited number of data points, a tentative maturity map for rocks of latest Cretaceous-Early Tertiary age can be drawn in VR equivalent for the northern subalpine massifs (Fig. 6).

No T_{max} measurement is available for sample 11 in the Swiss Alps, but the state of thermal degradation of the rare recognizable palynomorph fragments (Plate 1, Figs. 12 and 13) puts the thermal maturity level clearly in the gas zone (i.e. VR > 1.35%).

Discussion

Evidence of diagenetic/metamorphic transformations in the external part of the Western Alps has been derived from mineral assemblages and illite cristallinity by various authors: MARTINI & VUAGNAT (1965), MARTINI (1972), KÜBLER et al. (1974) and SAWATZKI (1975) have demonstrated that, in the Lower Tertiary graywackes of the Taveyannaz Formation, this metamorphism increases from the northern subalpine massifs in Haute-Savoie towards the Swiss Alps in the northeast and, in a given sector, towards the inner margin of the Helvetic domain, i.e. towards the external cristalline massifs. APRAHAMIAN et al. (1975) in the Platé Massif and KÜBLER et al. (1979) in the Bornes, Aravis and Platé Massifs (Fig. 1) made similar observations on various lithological units. The latter authors mainly studied the evolution of OM (vitrinite reflectance): although the geographical and stratigraphical spread of their data base makes detailed comparisons with our data difficult, VR measurements in the Bornes Massif clearly show an increase in thermal maturity towards the east. A similar gradient in OM maturity can be guessed from some Tertiary samples in the Bornes Massif (GORIN et al. 1989). Finally, clay minerals assemblages show an increasing diagenesis/metamorphism towards the east in the Ultrahelvetic units of the Sulens klippe (DECONINCK & CHAROLLAIS 1986).

Most of these authors associate the increase in metamorphism with the overthrusting of the Ultrahelvetic and Prealpine nappes, which resulted in the burial and a related increase in pressure and temperature of sediments deposited in the Helvetic domain. Increased metamorphism towards Central Switzerland in the northeast may be explained by a similar thickening of the piles of nappes, or by a steeper geothermal gradient in that direction.

It is well accepted that OM maturation in sedimentary rocks is mainly controlled by temperature and time and little by pressure (KÜBLER et al. 1979, CRUMIERE et al. 1988). Temperature, and consequently OM maturation, are a direct function of depth and heat flow. The NNE-SSW orientation of the tentative isomaturity lines on Fig. 6 seems subparallel to the original front of Ultrahelvetic and Prealpine nappes, which can be guessed from their present position north of the Arve River and from that of the Annes and Sulens klippes south of the River. The eastward maturity increase does not seem to be closely related to the proximity of external cristalline massifs: samples in the south of the study area (e.g. 1 to 7) appear less mature but lie closer to cristalline basement than samples in the north (e.g. 6686 and 6687). Consequently, part of the observed maturity gradient may be associated with the increased nappes overburden towards the east, prior to their erosion. This interpretation is in agreement with that of previous authors. If we use Lopatin's method for OM maturation (WAPLES 1980) and assume a normal thermal regime in the Tertiary (i.e. a geothermal gradient of 30 °C/km), our maturity estimates for samples 8, 9 and 10 (Fig. 6) give a range of burial depths in line with the 3 to 6 km figures given by SAWATZKI (1975) and KÜBLER et al. (1974) for the Lower Tertiary Taveyannaz Formation in this area.

As for sample 11, it is too isolated to draw conclusions from its very high thermal maturity, but a large part of its thermal evolution is probably associated with very thick nappe overburden.

Biostratigraphy of dinoflagellate cysts

Systematics

Sample 1 yielded a rich and well-preserved association of dinoflagellate cysts (Fig. 4; Plate 2). A preliminary, non-exhaustive study leads to the recognition of ca. 30 species.

a. Dinoflagellate cysts

- *Achomosphaera ramulifera* (Deflandre 1937b) Evitt 1963.
- *Areoligera* cf. *A. coronata* (O. Wetzel 1933b) Lejeune-Carpentier 1938.
- Areoligera cf. *A. tenuicapillata* (O. Wetzel 1933b) Lejeune-Carpentier 1938: Pl. 2, Fig. 8.
- *Areoligera* spp.
- *Bosedinia laevigata* (Jiabo 1978) He Cheng-quan 1984b.
- *Cerodinium boloniense* (Riegel 1974) Lentin & Williams 1989: Pl. 2, Fig. 3.
- *Cerodinium diebelii* (Alberti 1959b) Lentin & Williams 1987: Pl. 2, Fig. 4.
- *Cerodinium medcalfii* (Stover 1974) Lentin & Williams 1987.
- *Cordosphaeridium commune* Corradini 1973.
- *Cordosphaeridium fibrospinosum* Davey & Williams 1966b.
- *Coronifera oceanica* Cookson & Eisenack 1958 emend. May 1980.
- *Dinogymnium pustulicostatum* May 1977.
- *Dinogymnium westralium* (Cookson & Eisenack 1958) Evitt et al. 1967: Pl. 2, Fig. 10.
- *Exochosphaeridium phragmites* Davey et al. 1966.
- *Exochosphaeridium* spp.
- *Florentinia deanei* (Davey & Williams 1966b) Davey & Verdier 1973.
- *Hystrichosphaeridium tubiferum* (Ehrenberg 1838) Deflandre 1937b emend. Davey & Williams 1966b.
- *Hystrichosphaeridium* spp.
- *Hystrichostrogylon* sp. A.
- *Isabelidinium cooksoniae* (Alberti 1959b) Lentin & Williams 1977a: Pl. 2, Fig. 1.
- *Palaeocystodinium benjaminii* Drugg 1967.
- *Palaeohystrichophora infusorioides* Deflandre 1935: Pl. 2, Fig. 7.
- *Palaeotetradinium silicorum* Deflandre 1936b emend. Deflandre & Sarjeant 1970: Pl. 2, Fig. 13.
- *Phelodinium magnificum* (Stanley 1965) Stover & Evitt 1978: Pl. 2, Fig. 2.
- *Pierceites pentagona* (May 1980) Habib & Drugg 1987: Pl. 2, Fig. 11.
- *Raphidodinium fucatum* Deflandre 1936b emend. Sarjeant & Downie 1982: Pl. 2, Fig. 9.
- *Rigaudella apenninica* (Corradini 1973) Below 1982b.
- *Spiniferites membranaceus* (Rossignol 1964) Sarjeant 1970.
- *Spiniferites ramosus granosus* (Davey & Williams 1966a) Lentin & Williams 1973.
- *Spiniferites* spp.
- *Spongodinium delitiense* (Ehrenberg 1838) Deflandre 1936b emend. Lucas-Clark 1987: Pl. 2, Fig. 6.
- *Stephodinium coronatum* Deflandre 1936a: Pl. 2, Fig. 5.
- *Trigonopyxidia ginella* (Cookson & Eisenack 1960a) Downie & Sarjeant 1965: Pl. 2, Fig. 12.
- *Wallodinium anglicum* (Cookson & Hughes 1964) Lentin & Williams 1973.

b. Incertae sedis

- *Linotolypa* spp.

Palynostratigraphy

The dinocyst association encountered in sample 1 was compared with those already described from sediments of Late Cretaceous-Early Tertiary age in the following locations:

- Maastrichtian stratotype in the Netherlands: WILSON (1971 and 1974); HERNGREEN et al. (1986).
- Maastrichtian and Danian of Scandinavia: MORGENROTH (1968), KJELLSTRÖM (1973), HANSEN (1977 and 1979), KJELLSTRÖM & HANSEN (1981), HULTBERG (1985 and 1986), HANSEN et al. (1986b).
- France: JAN DU CHÊNE et al. (1975).
- Spain: DE CONINCK & SMIT (1982).
- Switzerland: VAN STUIJVENBERG et al. (1976), JAN DU CHÊNE (1977a and 1977b).
- Tunisia: BRINKHUIS & LEEREVELD (1988).
- USA: DRUGG (1967), BENSON (1976), MAY (1980), FIRTH (1987).
- India: JAIN et al. (1975).

These comparisons lead to the following observations:

1) The absence of species belonging to the genera *Odontochitina* DEFLANDRE 1935 and *Xenascus* COOKSON & EISENACK 1969 suggests an age younger than the boundary Early-Late Maastrichtian.
2) The presence of genera not extending beyond the Cretaceous-Tertiary boundary, such as *Dinogymnium* spp. and *Exochosphaeridium* spp., together with the absence of Danian marker species, particularly of *Danea mutabilis (=D. californica),* precludes a Palaeocene age.
3) The following association: *Cerodinium boloniense, Cerodinium diebelii, Cordosphaeridium fibrospinosum, Dinogymnium westralium, Isabelidinium cooksoniae, Palaeocystodinium benjaminii, Phelodinium magnificum, Pierceites pentagona, Spongodinium delitiense* and *Trigonopyxida ginella* enables us to date sample 1 as Late Maastrichtian.
4) The following species: *Palaeohystrichophora infusorioides* (Albian-Campanian), *Palaeotetradinium silicorum* (Turonian-Campanian) and *Stephodinium coronatum* (Albian-Campanian) indicate reworking of possible Turonian to Campanian age.

Discussion

The association of dinoflagellate cysts in sample 1 indicates a Late Maastrichtian age. This is in agreement with datings obtained by VILLARS (1988) from planktonic foraminifera and nannofossils.

This sample contains some reworked dinoflagellate cysts of Turonian-Campanian age. VILLARS (1988) has shown that there is a major erosional unconformity surface at the base of the Wang Formation, which erodes the underlying Senonian marls and limestones and locally cuts down to the Lower Cretaceous deposits. It is therefore not surprising to find reworked material in the lower part of the Wang Formation: sample 1 comes from the Col de Bellefond section where only the lowermost part of the Wang Formation has been preserved from Early Tertiary erosion (VILLARS 1988).

Conclusions

Organic matter (OM) of the Wang Formation was studied in the northern subalpine massifs using geochemical (Roch-Eval pyrolysis) and palynological methods. The organic content of the ten analysed samples gives some informations on the palaeoenvironment, thermal history and age of the deposits.

The organic content of the Wang Formation is low (0–0.9% TOC). It is dominated by type III OM (i.e. humic fragments of the inertinite and vitrinite groups) mixed with a variable amount of dinoflagellate cysts and cutinite. This palynofacies characterizes open marine, well-oxygenated depositional conditions compatible with those of an outer hemipelagic carbonate shelf (VILLARS 1988). The high sedimentation rate calculated by this author is supported by the dilution of OM and the excellent preservation state of all organic particles (in the thermally immature samples).

Thermal maturity of the Wang Formation was tentatively established from Rock-Eval T_{max} measurements and palynomorph thermal alteration in natural and UV light. Maturity increases from west to east, as observed previously in Tertiary samples from the Bornes Massif. Tentative isomaturity lines, valid in the northern subalpine massifs for Upper Cretaceous and Lower Tertiary sediments deposited in the Helvetic domain, seem to run in a NNE-SSW direction, subparallel to the presumed original front of the Ultrahelvetic and Prealpine nappes (now evidenced only by klippes in the Bornes Massif). These observations support previous interpretations associating this maturity gradient with the burial of the Helvetic domain under the Ultrahelvetic and Prealpine nappes. Nevertheless one cannot totally exclude the overprint of a steepening geothermal gradient towards the northeast.

One of the immature samples in the Chartreuse Massif has yielded a rich assemblage of dinoflagellate cysts. Comparisons with existing studies on Late Cretaceous-Early Tertiary cysts indicate a Late Maastrichtian age. It is a confirmation of age datings obtained by VILLARS (1988) from planktonic foraminifera and nannofossils.

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Plate 1

Palynofacies of the Wang Formation in the northern subalpine massifs.
Figs. 1 to 3 at scale A, Figs. 4 to 13 at scale B.

Figs. 1, 2 and 3. General view of palynofacies at three different locations with different thermal maturity level.

Fig. 1. Sample 1, slide FV1440c. Dark-coloured to black constituents are mainly vitrinite and inertinite fragments. Note their small size and often subrounded shape. Light-coloured constituents are palynomorphs, essentially dinoflagellate cysts. Sample thermally immature (TAI = 1–2).

Fig. 2. Sample 6, slide FV1340. Black to dark-coloured inertinite, vitrinite and cutinite fragments are larger and more angular than on Fig. 1. Palynomorphs are essentially dinoflagellate cysts. Thermal maturity is close to or just reaching the oil zone (TAI = 2–2.5).

Fig. 3. Sample 9, slide FV780. Size and angularity of black vitrinite and inertinite fragments are comparable with those in sample 6. Palynomorphs (= a) are darkened by thermal maturity, in the late stage of the oil zone (TAI = 3–3.5).

Figs. 4 to 13. Show details of palynofacies and illustrate the increasing darkening and deterioration of palynomorphs with increasing thermal maturity.

Fig. 4. Sample 1, slide FV1440c. Thermally immature (TAI = 1–2).
b = dinoflagellate cysts
c = inertinite + vitrinite

Figs. 5, 6. Sample 7, slide FV1000e. Thermal maturity is close to or just reaching the oil zone (TAI = 2–2.5).
d = dinoflagellate cysts
e = inertinite + vitrinite

Fig. 7. Sample 5, slide FV1337a. Thermal maturity is close to or just reaching the oil zone (TAI = 2–2.5).
f = dinoflagellate cysts
g = inertinite + vitrinite

Fig. 8. Sample 6, slide FV1340. Thermal maturity is close to or just reaching the oil zone.

Fig. 9. Sample 8, slide FV655b. Thermal maturity is in the early stage of the oil zone (TAI = 2.5–3).
h = dinoflagellate cyst darkened by thermal maturity
i = inertinite

Figs. 10, 11. Sample 9, slide FV780a. Thermal maturity is in the late stage of the oil zone (TAI = 3–3.5).
j = thermally darkened and degraded dinoflagellate cysts/palynomorphs
k = inertinite

Figs. 12, 13. Sample 11, slide FV1448a. Wang Formation in the Wildhorn nappe of the Swiss Alps. Thermal maturity is in the gas zone (TAI = 3.7–4).
l = fragment of thermally degraded dinoflagellate cyst
m = ghost of thermally degraded bisaccate pollen
n = inertinite
o = "amorphous organic matter" originating from the thermal degradation of palynomorphs

Plate 1

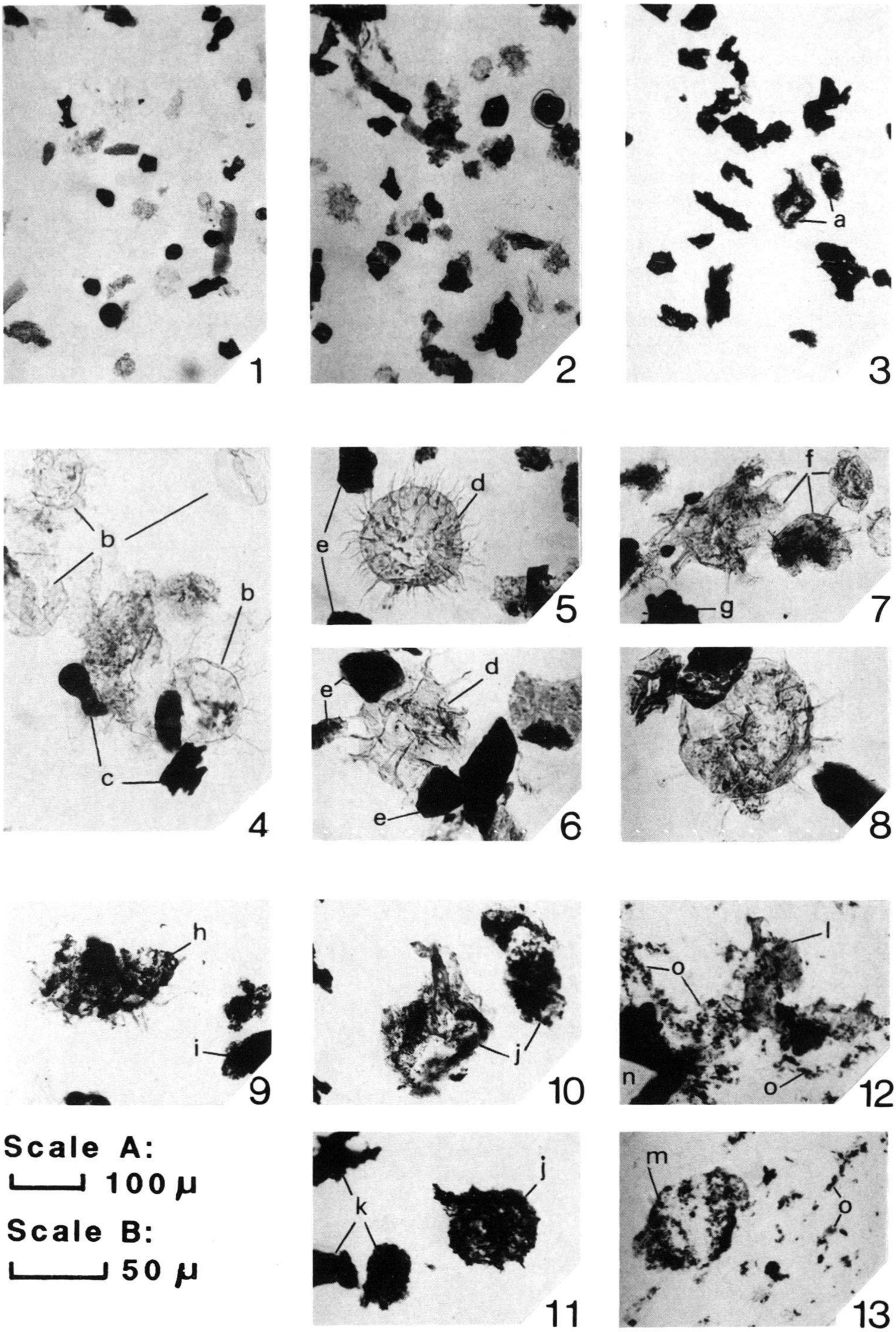
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Plate 2

Dinoflagellate cysts of the thermally immature sample 1 in the Chartreuse Massif. All photographs taken in Nomarski interference contrast, LF=low focus, MF=medium focus, HF=high focus. G45/2, etc.="England Finder" coordinates.
Figs. 5, 7 and 13 are reworked cysts of possible Turonian to Campanian age.

Fig. 1. *Isabelidinium cooksoniae* (ALBERTI 1959b) LENTIN & WILLIAMS 1977a, LF, dorsal view, slide FV1440e, T40/4.

Fig. 2. *Phelodinium magnificum* (STANLEY 1965) STOVER & EVITT 1978, LF, ventral view, slide FV1440g, U31/2.

Fig. 3. *Cerodinium boloniense* (RIEGEL 1974) LENTIN & WILLIAMS 1989, HF, dorsal view, slide FV1440e, V33.

Fig. 4. *Cerodinium diebelii* (ALBERTI 1959b) LENTIN & WILLIAMS 1987, HF, right lateral view, slide FV1440g, Z49/1.

Fig. 5. *Stephodinium coronatum* DEFLANDRE 1936a, HF, slide FV1440f, G37/4.

Fig. 6. *Spongodinium delitiense* (EHRENBERG 1838) DEFLANDRE 1936b emend. LUCAS-CLARK 1987, MF, slide FV1440e, S20/2.

Fig. 7. *Palaeohystrichophora infusorioides* DEFLANDRE 1935, HF, slide FV1440e, Q43/3.

Fig. 8. *Areoligera* cf. *A. tenuicapillata* (O. WETZEL 1933b) LEJEUNE-CARPENTIER 1938, MF, slide FV1440f, Q24/4.

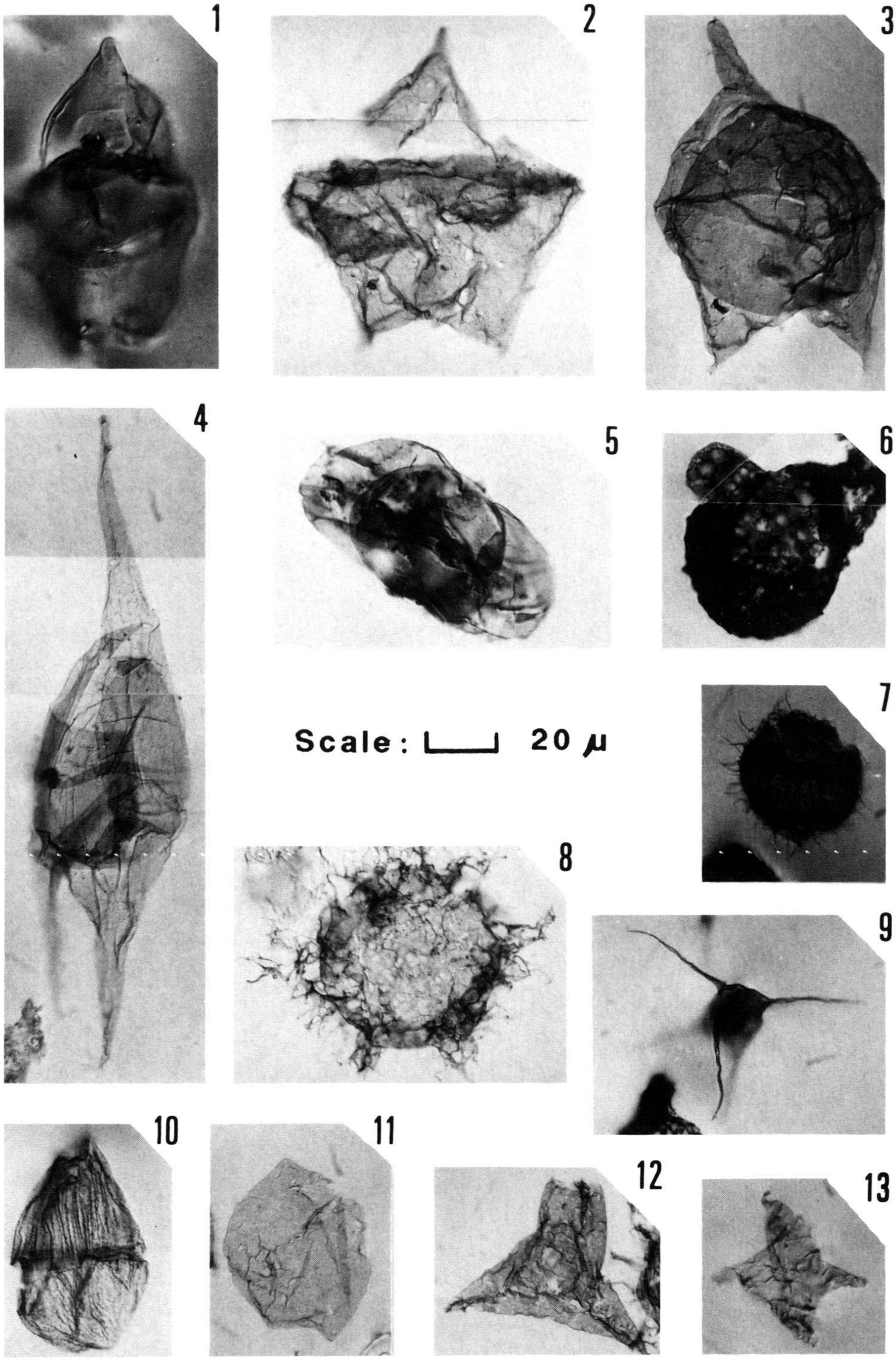
Fig. 9. *Raphidodinium fucatum* DEFLANDRE 1936b emend. SARJEANT & DOWNIE 1982, HF, slide FV1440b, M32/3.

Fig. 10. *Dinogymnium westralium* (COOKSON & EISENACK 1958) EVITT et al. 1967, HF, ventral view, slide FV1440b, P28/4.

Fig. 11. *Pierceites pentagona* (MAY 1980) HABIB & DRUGG 1987, MF, slide FV 1440f, D37.

Fig. 12. *Trigonopyxidia ginella* (COOKSON & EISENACK 1960a) DOWNIE & SARJEANT 1965), MF, slide FV1440b, J22/1.

Fig. 13. *Palaeotetradinium silicorum* DEFLANDRE 1936b emend. DEFLANDRE & SARJEANT 1970, HF, slide FV1440b, R30/1.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