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Sedimentology of the Upper Marine Molasse of the Rhône-Alp Region, Eastern France: Implications for Basin Evolution

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ABSTRACT

The Voiron-Chambéry-Annecy region (Rhône-Alp) is situated in an area of interfingering and juxtaposed tectonic and stratigraphic styles. It was affected in the Palaeogene by extensional tectonics associated with the opening of the Rhine-Bresse-Rhône system. In the Neogene it was overwhelmed by flexural subsidence in a peripheral foredeep related to the WNW-wards advance of the Alpine orogenic wedge. Late Miocene and Pliocene deformation has resulted in the formation of tight folds and thrust faults along the N–S trend of earlier Palaeogene (and even Mesozoic) extensional lineaments, incorporating the region into the southern prolongation of the Jura structural domain.

Up to 1150 m of Miocene marine sediment is preserved in the basin. The basin-fill has been divided into five lithosomes plus a basal transgressive unit. Initial flooding in the latest Aquitanian to early Burdigalian from the south established a vigorous tidal strait occupied by large scale subtidal sandwaves. A tide-dominated coast prograded into the strait from the west (Tresserve and Forezan lithosomes). By the end of the Burdigalian there was a basin-wide interlude of low energy muddy shelf conditions (Montaugier lithosome). The Langhian to Serravallian was characterized by the progradation of a fine-grained tide-dominated coast (Grésy lithosome) and then the advance of thick sandy and gravelly wedges (Pont-de-Beauvoisin lithosome) from the Alpine flank of the basin into a tidal sea, eventually occluding it from the Alpine perimeter.

The Miocene sediments can be divided into two stratigraphic packages, each with a surface of marine onlap at its base. The first sequence (early to late Burdigalian) is restricted relatively to the east. It is marked at its top by stratigraphic offlap corresponding to deposition of the Montaugier lithosome. The second stratigraphic sequence (early Langhian to Serravallian) coarsens up and strongly onlaps the western basin margin, spreading into the southern Bresse graben to the margin of the Massif Central.

RÉSUMÉ

La région Voiron-Chambéry-Annecy (Rhône-Alp) est caractérisée par l'imbrication et la juxtaposition de types tectoniques et stratigraphiques différents. Au cours du Paléogène elle est affecté par la tectonique distensive liée à l'ouverture du système Rhin-Bresse-Rhône, alors qu'au Néogène elle est dominée par la flexion crustale due à l'avancement du bâti alpin, en direction du WNW, sur le bassin d'avant-pays. Les déformations au cours du Miocène supérieur et du Pliocène conduisent à la genèse de plis serrés et de chevauchements qui peuvent reprendre la trace des accidents distensifs du Paléogène inférieur ou même du Mésozoïque; le domaine se trouve alors intégré dans la partie méridionale de la chaîne du Jura.

Jusqu'à 1150 m d'épaisseur de sédiments marins du Miocène se trouvent accumulés dans le bassin. Cette série peut être subdivisée en cinq «lithosomes» superposés à une unité transgressive basale: la montée marin initiale à partir du Sud au cours de l'Aquitanien supérieur – Burdigalien inférieur crée un détroit dominé par des marées vigoureuses, et occupé par des mégarides de sable subtidales. Une ligne côtière prograde dans ce détroit à partir de l'ouest (lithosomes de Tresserve et de Forezan). Vers la fin du Burdigalien, l'ensemble du bassin passe, temporairement à des conditions de basse énergie à sédimentation pélitique (lithosome de Montaugier). Du Langhien au Serravallien, la sédimentation côtière fine, dominée par l'action des marées (lithosome de Grésy), puis des cônes sablo-graveleux (lithosome de Pont-de-Beauvoisin) progressent de la bordure alpine vers la bassin marin, probablement en s'écartant de la bordure alpine.

Les dépôts d'âge miocène peuvent être subdivisés en deux pacquets stratigraphiques, chacune limitée à sa base par un biseau d'aggradation («onlap») marin. La première pacquet (ou séquence) (Burdigalien inférieur à supérieur) est limitée au secteur oriental. Elle est marquée à sa toit par une unité de progradation («offlap») correspondant au dépôt du lithosome de Montaugier. La deuxième pacquet (Langhien inférieur à Serravallien) montre une granocroissance vers le haut, et des biseau d'aggradation très marqués en direction du bord Ouest du bassin; elle se développe jusque dans les Sud du fossé de la Bresse et aux abords du Massif Central.

ZUSAMMENFASSUNG

Die Region zwischen Voiron, Chambéry und Annecy (Rhône-Alpen) zeichnet sich durch Verzahnung nebeneinanderliegender und unterschiedlicher tektonischer Formen und stratigraphischer Abfolgen aus. Im Paläogen wurde das Gebiet durch Zerrungstektonik beeinflusst, ausgelöst durch die Öffnung des Rhein-Bresse-Rhône Systems. Später im Neogen dominiert Subsidenz einer randlichen Vortiefe als Folge der Lithosphärenbelastung durch das Vorwandern des alpinen Deckenstapels in Richtung WNW. Die Spät-Miozäne und Pliozäne Deformation bewirkte die Bildung enger Falten und Überschiebungen entlang älterer, paläogener (oder sogar mesozoischer) N–S streichender Dehnungsbrüche. Dabei wird das Gebiet in die strukturelle Entwicklung des südlichen Teils der Jura-Kette einbezogen.

Gegen 1150 m mariner miozäner Sedimente sind im Becken erhalten. Diese Beckenfüllung wurde in fünf Lithosome unterteilt. Dazu kommt noch eine basale, transgressive Einheit. Die initiale Flutung zwischen dem spätesten Aquitanian und dem frühen Burdigalian von Süden her führte zu einer strömungsreichen, gezeitendominierten Meeresstrasse, wobei sich grosse subtidale Sandwellen bildeten. Eine gezeitendominierte Küste progradierte von Westen her gegen diese Meerenge (Lithosome von Tresserve und Forezan). Gegen Ende des Burdigalian begann eine Episode mit Bedingungen geringer Energie, was die Ablagerung schlammiger Schelf-Sedimente mit sich brachte (Montaugier Lithosom). Im Langhian und Serravallian progradierten feinkörnige, gezeitendominierte Küstensedimente (Grésy Lithosom), bevor sandige und geröllführende Sedimentkeile (Pont-de-Beauvoisin Lithosom) gebildet wurden, deren Material von den Flanken des alpinen Orogens stammt. Dies trennte vermutlich das alpine Randmeer vom offenen Gezeitenmeer.

Die miozänen Ablagerungen können in zwei grössere Einheiten unterteilt werden, wobei jede an ihrer Basis einen marinen Onlap zeigt. Die erste Sequenz (frühes bis spätes Burdigalian) ist eher auf den östlichen Teil beschränkt. Das Dach dieser Einheit ist durch einen stratigraphischen Offlap markiert, gekennzeichnet durch die Ablagerung des Montaugier Lithosom. Die zweite Sequenz (frühes Langhian bis Serravallian) zeigt Kornvergröberung gegen oben und starken Onlap gegen den westlichen Beckenrand. Sie entwickelt sich im südlichen Bresse-Graben und bis hin zum Rand des Massif Central.

1. Introduction

The Voiron-Chambéry-Annecy area of the Rhône-Alp region of eastern France lies within the Alpine peripheral foreland basin, flanked to the east by the frontal (Subalpine) chains of the Alpine orogen such as the Bauges and Chartreuse, and to the west by the north-south folds of the Jura province and the sedimentary trough of the Bresse-Rhône graben in the Bas-Dauphiné. It is therefore a pivotal area between the extensional western European rift system (Rhine-Bresse-Rhône system) and the flexurally-driven peripheral foreland basin system (Figs. 1, 2). It is also a linking area between better-documented parts of the foreland basin to the south of Grenoble (Haug 1891; Goguel 1936; De Lapparent 1938; Goguel 1948; Gigot et al. 1974; Beaudoin et al. 1975; Elliott et al. 1985) and to the northeast of the Lake of Geneva (Matter et al. 1980; Allen et al. 1985; Homewood et al. 1986).

The 1150 m-thick shallow marine deposits of the Upper Marine Molasse (OMM – from *Obere Meeresmolasse*) form part of the Tertiary fill of the basin, generally lying unconformably on thick Mesozoic strata. The OMM in this area ranges from early to middle Miocene in age (Burdigalian-Serravallian), and was deposited in a seaway that extended from the Mediterranean region in the south, around the arcuate Alpine perimeter to the Swiss Molasse basin and beyond to the east (Fig. 3). This seaway, though extensive around the arc, was restricted in its width. No doubt the funnel-like shape of the marine basin, wide in the Rhodano-provençal region in the south and narrow in the Rhône-Alp region in the north, was influential in allowing the generation of the high tidal ranges in the seaway interpreted from sedimentary facies and successions.

Previous workers have largely concentrated on the stratigraphy and petrography of the marine molasse (Douxami 1899; Revil & Roch 1925; Vatan 1949; Vatan et al. 1957; Michel & Caillon 1957; Gidon 1960, 1969a, b, 1970; Latreille 1969; Gigout 1969; Donze 1972; Lamiraux 1977; Mujito 1981; Meylan 1982; Rigassi 1982; Berger 1985; Perriaux 1984; Perriaux et al. 1984). Other workers have established the structural geometry of the

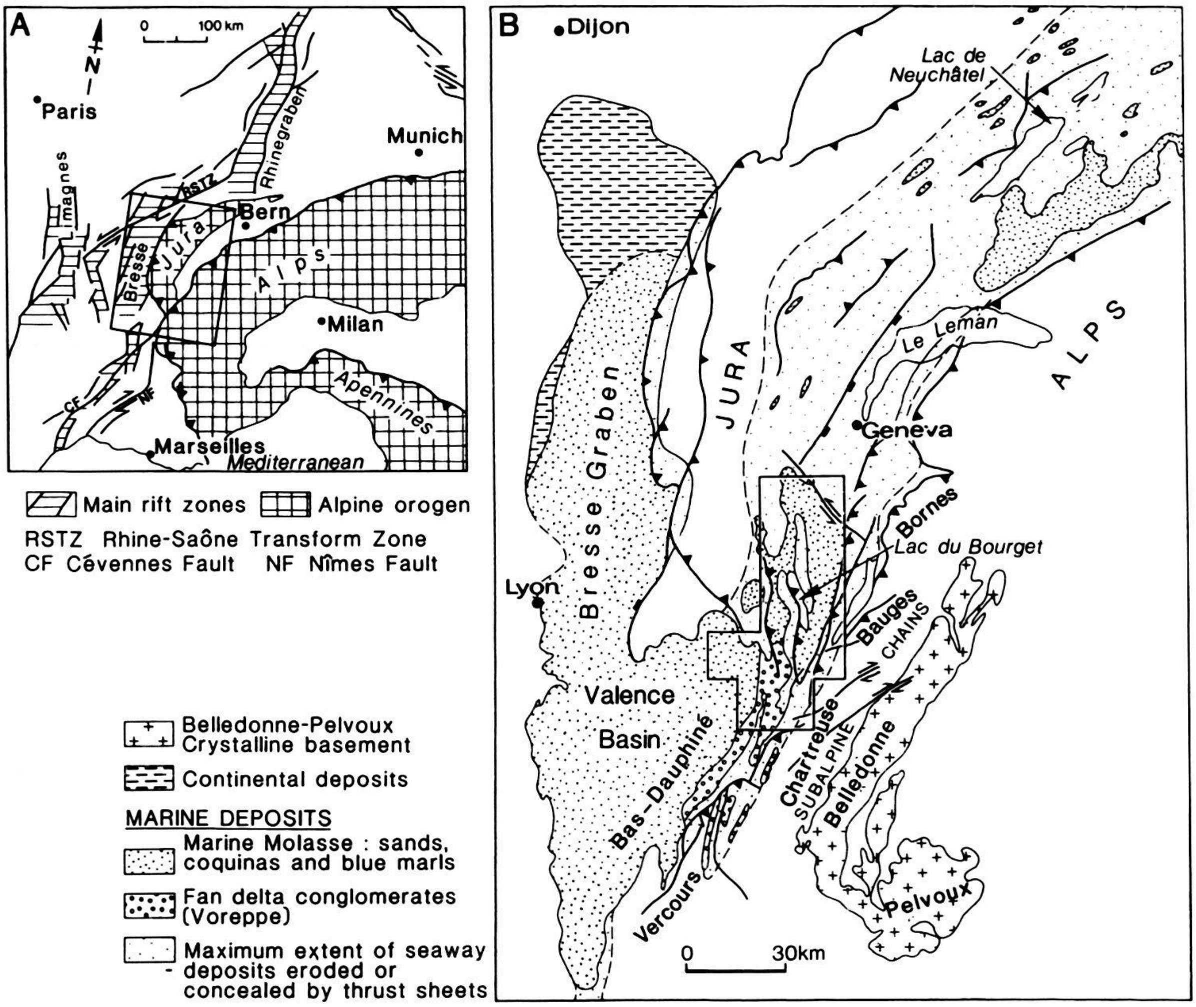


Fig. 1. **A** The western European rift system and Alpine orogenic belts (after Bergerat et al. 1990). **B** The Bresse-Rhone graben, Jura, Molasse Basin and Alpine arc in eastern France and Switzerland (after Debrand-Passard & Courbouleix 1984). The Belledonne-Pelvoux crystalline bsasement massif is shown for reference – other crystalline massifs are not shown.

area, dominated by the north-south aligned tight folds of the southern prolongation of the Jura, and the more strongly allochthonous units of the Subalpine chain (Santos-Narvaez 1980; Mugnier & Ménard 1986; Ménard 1988; Guellec et al. 1989, 1990; Butler 1989, 1991). The aim of this study was to provide a documentation of the marine facies present in the seaway and to make some palaeogeographical reconstructions for the Burdigalian-Serravallian time period. These data should prove useful in the future synthesis of the dynamics of the Miocene peri-Alpine seaway in France and Switzerland.

2. Stratigraphy

The rocks of the Rhône-Alp region range in age from Mesozoic to Miocene. The OMM was deposited during the time period represented by the Burdigalian, Langhian and Serravallian stages of the Miocene. These Miocene marine sediments have a sharp

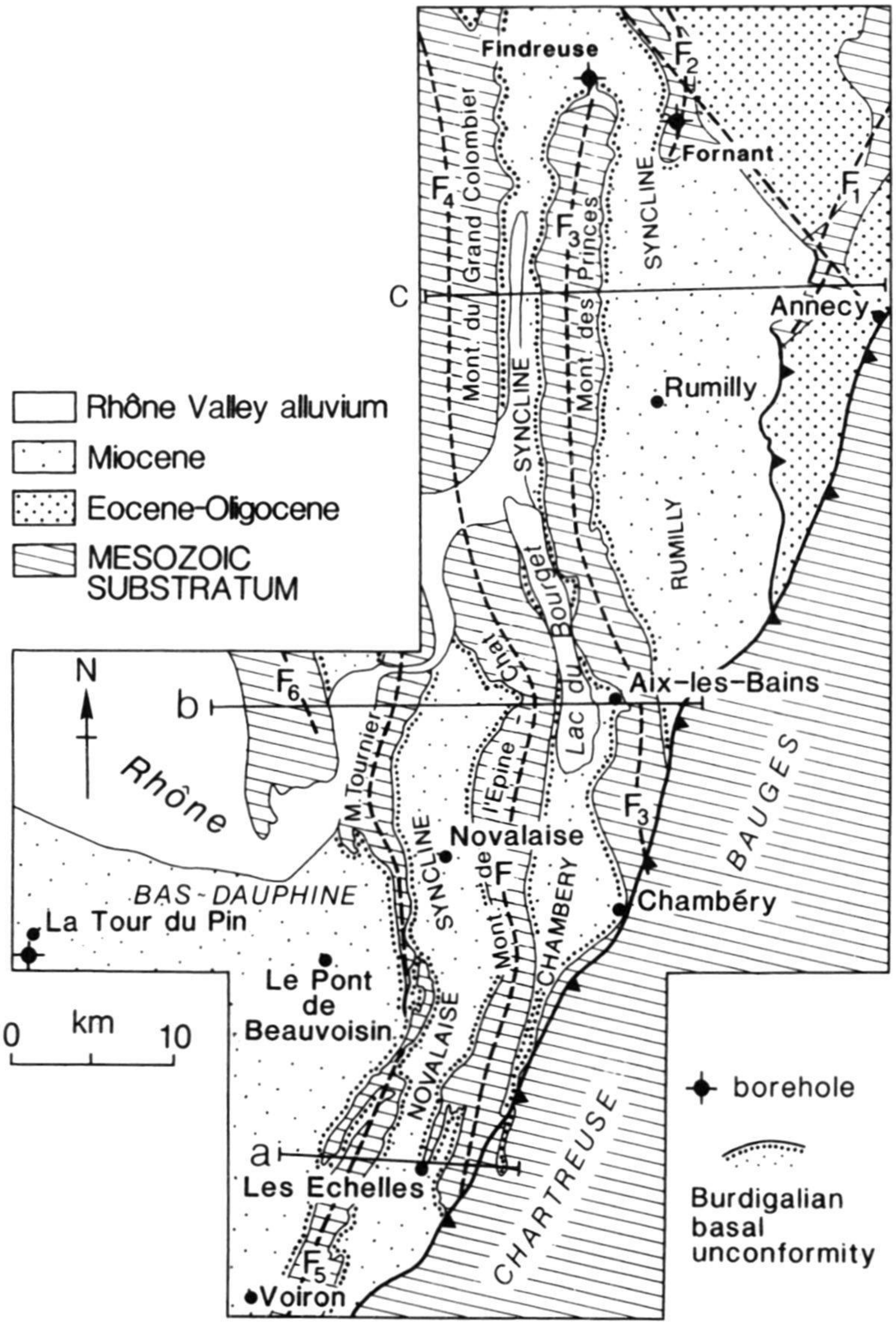


Fig. 2. Geological map of the Rhoñe-Alp region showing the N–S anticlines of Mesozoic rocks and adjacent synclines containing Molasse. Cross-sections a, b, c are shown in Figure 5. The locations of the boreholes at Fornant and Findreuse (Savoie) and close to La Tour du Pin (Bas-Dauphine) are shown.

boundary with underlying Lower Freshwater Molasse, as, for example in the north of the study area in the region of Rumilly and Seyssel, which has been dated as latest Aquitanian or basal Burdigalian using macrofossil and microfossil evidence cited in Berger (1985). The youngest Lower Freshwater Molasse directly beneath the OMM in the area of the two boreholes at Fornant and Findreuse (Fig. 2) has been dated using a magne-

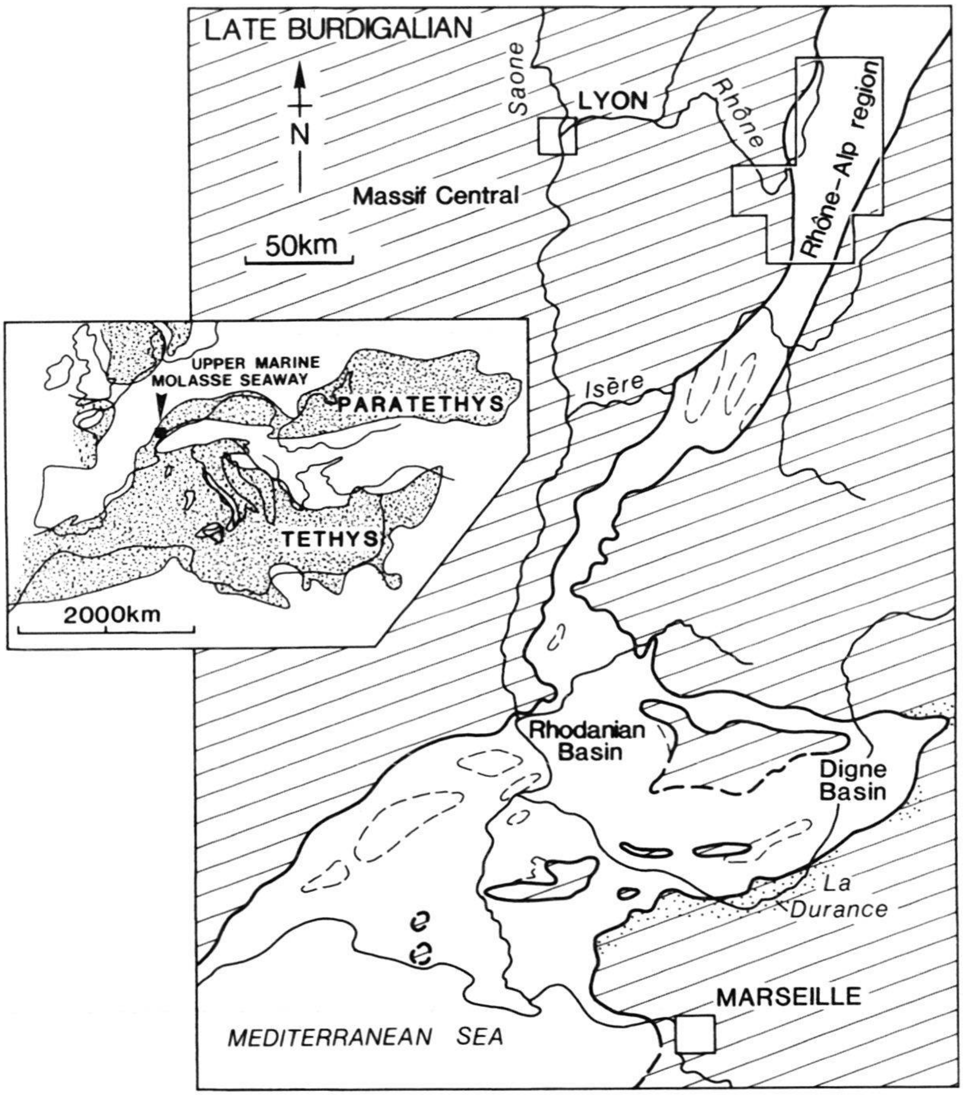


Fig. 3. The palaeogeographical reconstruction of the OMM seaway during the Burdigalian (compare with expanded seaway in Figure 1 B) (after Demarcq 1984). Eastern closure of Paratethys in inset is uncertain.

tostratigraphic chronology as < 21.5 Ma (Aquitanian) (Burbank, Engesser et al., 1992). The biostratigraphical assignment based on the Swiss mammal zones is from the La Chaux to Brüttelen 2 levels, as in the Mittelland of western Switzerland (Berger 1992). The base of the OMM is rarely dated from mammals. In western Switzerland, marine sediments already occur between the levels of La Chaux and Vully 1, that is, within the Aquitanian, so that no major stratigraphic gap between the Lower Freshwater Molasse and the Upper Marine Molasse is discernible in the plateau region of western Switzerland. In Savoie and the northern part of the Rhône-Alp region (Rumilly-Seyssel), the situation is a little less clear, but the base of the OMM may also be latest Aquitanian to early Burdigalian (Latreille 1969). This supports the view that little erosion has taken place at the boundary between the Lower Freshwater Molasse and the Upper Marine Molasse in the north of the study area. The basal OMM must have an age date close to 21 Ma. Elsewhere, particularly in the south of the Rhône-Alp study area, the OMM directly overlies Mesozoic strata with a large chronostratigraphic gap. In general, the biostratigraphical control *within* the OMM is very poor, though the top of the OMM can be dated by the first occurrence of planktic foraminifera such as *Orbulina universa* and *Orbulina suturalis*, demonstrating an age of N8 or younger (Langhian or Serravallian). Previous workers have been forced into making primarily a lithostratigraphical subdivision of the succession (Lamiraux 1977; Mujito 1981), dated where possible by poor microfaunal assemblages (Latreille 1969), or correlated loosely with similar lithological units outside of the region which have more reliable biostratigraphical assignments (Perriaux 1984).

The lithostratigraphical units defined by Lamiraux (1977) have, in places, interfingering relationships along their boundaries, so that parts of some units are age-equivalent to parts of other units. We therefore have built our stratigraphic framework on the basis of the work of Lamiraux (1977), differentiating five *lithosomes* – a term which Wheeler & Mallory (1956) used for "*a rock mass of essentially uniform or uniformly heterogeneous lithologic character, having intertonguing relationships in all directions with adjacent masses of different lithologic character*". The succession is well differentiated into these lithosomes in the southern half of the study area between Chambéry and the Bas-Dauphiné (Fig. 4), but the stratigraphy is less well differentiated towards the north of the area in Haute-Savoie. However, the Montaugier unit acts as a marker that is found throughout the region; it does not exhibit obvious interfingering relationships with other lithosomes.

3. Structure and Tectonic Setting

Closure of the Piemont/Tethyan ocean and collision of Adria and Europe (Tapponnier 1977) resulted in the shortening of the European margin and the downflexing of the European plate (Karner & Watts 1983; Mugnier & Ménard 1986; Homewood et al. 1986). The resulting foreland basin filled firstly with Eocene to lower Oligocene marine sediments typified by the North Helvetic Flysch of Switzerland and Haute Savoie (principally the Taveyannaz and Val d'Illiez Formations) and the Annot and Champsaur Formations of Haute Provence and les Hautes-Alpes. The basin then filled essentially to sea level during the Molasse phase (Oligocene to mid-Miocene). Telescoping of the European margin to the east of the study area in the Oligocene to late Miocene resulted in the deformation of these foreland basin sediments (principally the Eocene-early Oligocene flysch-like sediments and the Chattian-Aquitanian Lower Freshwater Molasse, together with, in the Chartreuse, the Burdigalian OMM).

Continued compression from the Alpine wedge since the end-Miocene caused folding of the Mesozoic substrate of the region together with its Tertiary cover, forming the NNE–SSW trending southern prolongation of the Folded Jura. The shortening accompanying this phase of deformation is, however, slight compared to the large displacements in the orogenic belt proper (Mugnier et al. 1987; Gratier et al. 1989). Section balancing indicates that the shortening across the Jura folds in the southern part of the study area is about 5%, with this value increasing progressively towards the north into the Jura fold-thrust belt (Chauve et al. 1988; Guellec et al. 1990). The western limbs of the Jura folds are commonly cut by steep thrust faults (Fig. 5). These may be older faults dating from the Rhine-Bresse-Rhône extensional phase that have been inverted, creating folds in their hangingwalls. The OMM is preserved in synclines between these Jura folds.

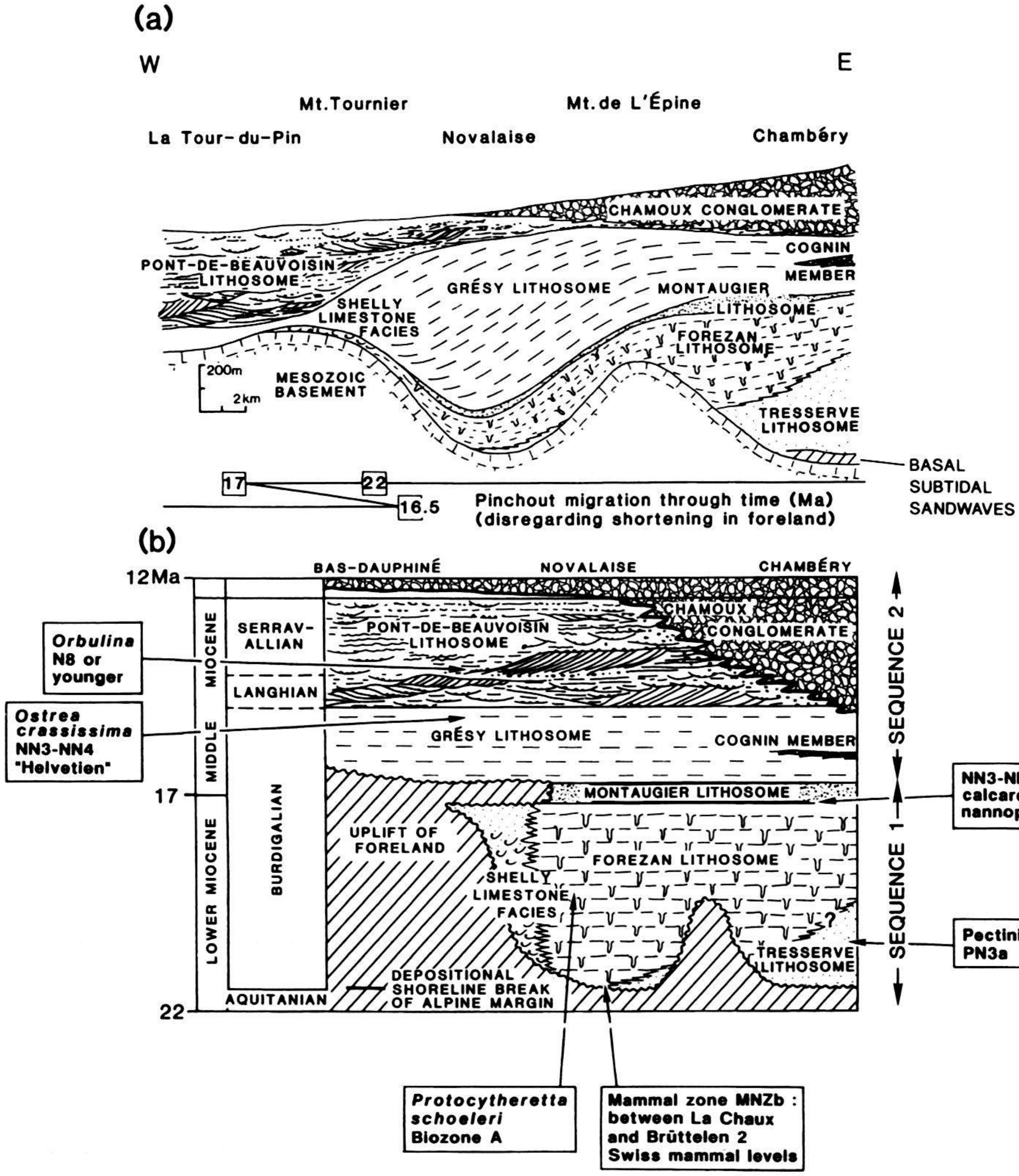


Fig. 4. Lithostratigraphic (a) and chronostratigraphic (b) cross-sections of the OMM in the south of the Rhône-Alp study area. Adapted from Lamiraux (1977), with approximate absolute ages derived largely from the European Oligocene-Miocene correlation chart of Berger (1992). Key faunal elements for chronostratigraphy are annotated. The lithosomes are named from villages nearby representative sections.

Although Eocene E–W folds and faults related to the compressional Pyrenean-provençal phase of deformation are found in the Rhône valley and Basse-Provence, strongly influencing molassic sedimentation (Gigot et al. 1974; Jones 1988), no such structures have been positively identified further to the north in the Rhône-Alp region (Siddans 1983).

During the late Eocene to Oligocene, the western European area underwent E–W extension, forming the linked rift basin system of the Rhine-Bresse-Rhône (Goguel 1948; Rat 1978; Ziegler 1988; Bergerat 1987; Bergerat et al. 1990). NNE–SSW trending

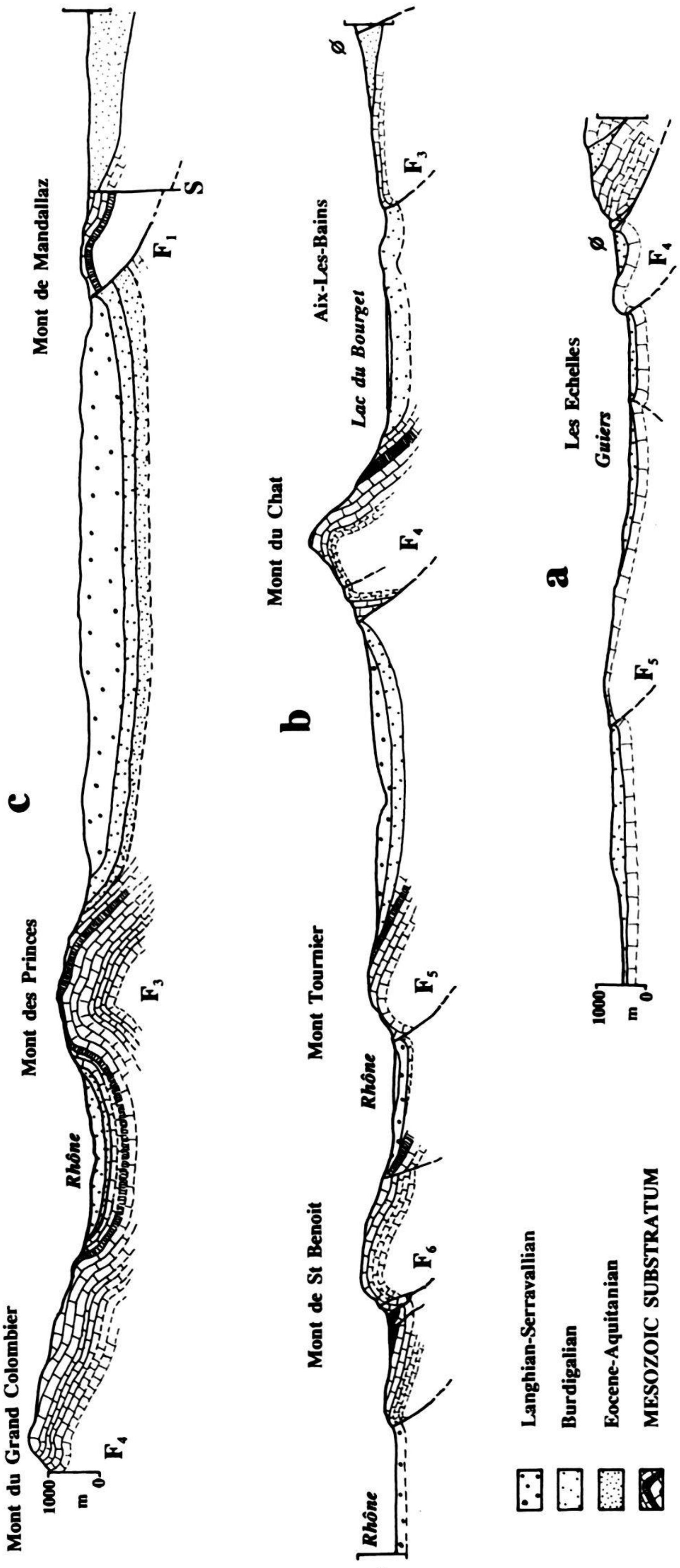


Fig. 5. Structural cross-sections across the Rhône-Alp region derived from geological maps. Sections a to c are located in Figure 2.

extensional faults of the Bresse graben and Jura province (Laubscher 1988; Chauve et al. 1988) extended southwards into the Rhône-Alp region and even into the Vercors and Chartreuse (Gidon 1982; Butler 1991, p 289), compartmentalizing the molasse basin during the Oligocene to early Miocene.

The OMM is found, therefore, in a number of structural positions: (i) within the Alpine orogenic wedge (basal deposits in the Chartreuse), (ii) within synclines between Jura folds (Rumilly, Chambéry and Novalaise synclines) and (iii) in the little deformed Bas-Dauphiné.

4. Sedimentology of the Upper Marine Molasse

4.1 The Initial Flooding

The Rhône-Alp region was flooded during the European micromammal biozone MN2b (Berger 1983, 1985), corresponding to the late Aquitanian-earliest Burdigalian (about 22 Ma). The marine transgression advanced along the peri-Alpine basin both from the SW (the *rhodanienne* origin of Rigassi 1977a; Büchi & Schlanke 1977; Berger 1985) and from the NE (the *viennoise* origin of Vavra 1982, and confirmed by Berger 1985). Working on sections in Haute-Savoie and western Switzerland, Berger (1985) concluded that the transgression was diachronous. In the case of Haute-Savoie, and therefore also the Rhône-Alp region, it originated from the south.

After the marine transgression, the Rhône-Alp region was occupied by a seaway that widened towards the NE into the Swiss Molasse Basin and to the SW into the "Rhodano-provençal Gulf" of Provence and the southern Rhône valley (De Lapparent 1938; Demarcq 1970, 1984) (Fig. 3). The western shore was rocky (Demarcq 1962; Latreille 1969), with Miocene marine sediments pinching out onto Cretaceous limestones (eg. at Gorges de Chailles; 8661 3582). The eastern shore has since been obscured by thrusting in the Subalpine chains and is thought to have extended further to the east to a position close to the base-Burdigalian uplifting thrust front (Demarcq 1962; Latreille 1969). The local presence of irregular unconformity surfaces on Cretaceous limestones (Fig. 6) covered with the traces of rock-boring bivalves, and limestone conglomerates and breccias containing bored pebbles, led Demarcq (1962), Perriaux (1984) and Perriaux et al. (1984) to conclude that the seaway at this time was characterized by numerous shallow shoals and islands.

The subcrop map of the unconformity (Fig. 7) shows that the topography of the surface was far from smooth, with wide variations in the amount of stratal omission below it. Actual amounts of missing stratigraphy are difficult to assess since lateral variations in Mesozoic-Palaeogene stratigraphy undoubtedly existed. However, it is evident that (1) progressively more substratum has been removed by erosion from north to south (north of Rumilly, the unconformity rests on Aquitanian, Chattian and Stampian deposits (Michel & Caillon 1957), whereas in the south the unconformity cuts down to Mesozoic carbonates, so that in the south of the study area (eg. at Les Echelles), in the Chartreuse and the Vercors, Miocene sediments rest directly on Lower Cretaceous Urgonian limestones); (2) there are local rapid changes in the stratigraphic level to which the unconformity cuts down, and positions of postulated palaeohighs are associated with distinctive facies within the Tertiary, such as the continental breccias of the Chattian-

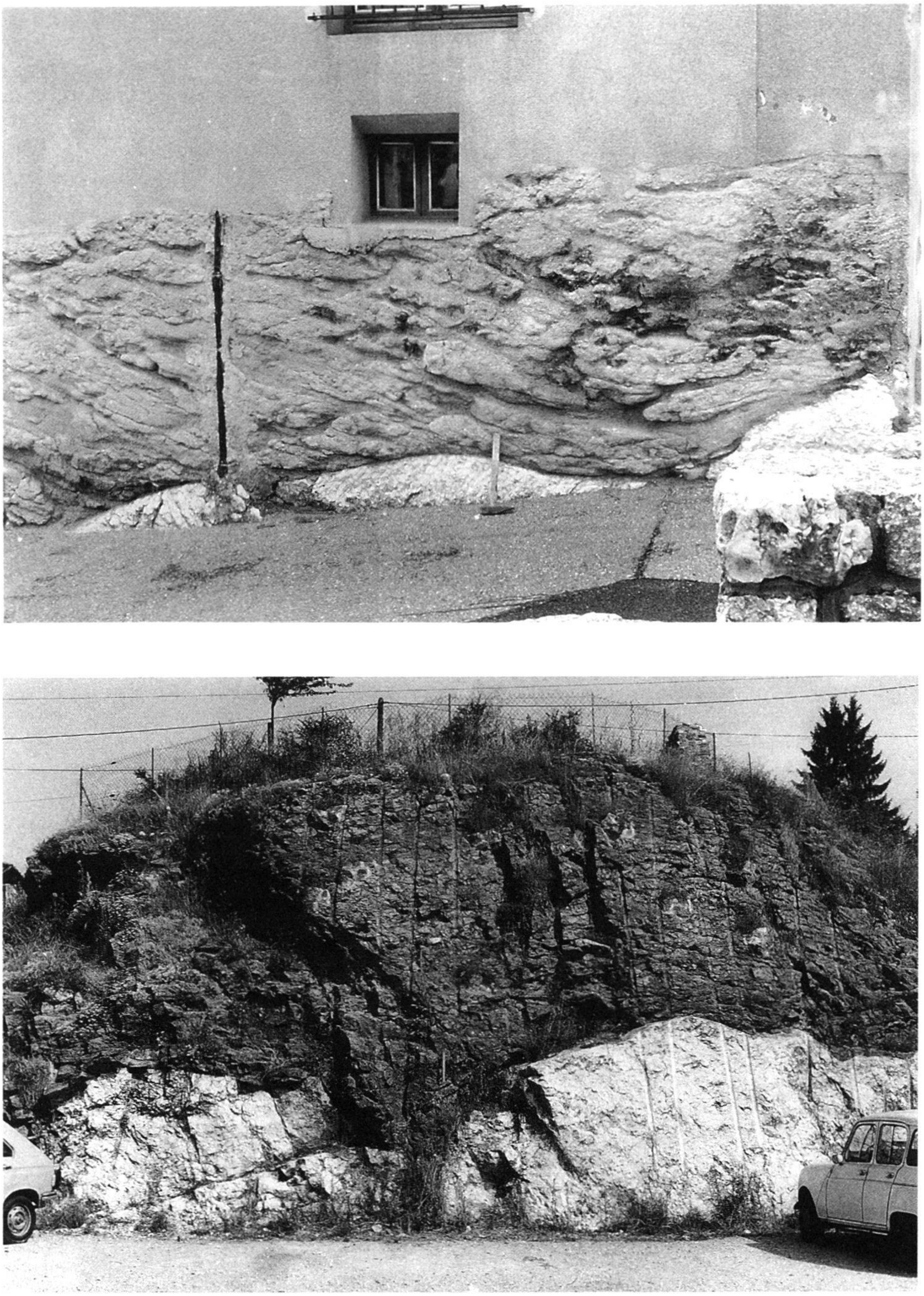


Fig. 6. The irregular unconformity surface of the Burdigalian Upper Marine Molasse resting on Cretaceous Urgonian limestones. Large-scale cross-strata of a subtidal sandwave overlie the unconformity at 867.5/354.1, in Les Echelles on the N6 road. Traces of rock-boring bivalves are found on the unconformity surface.

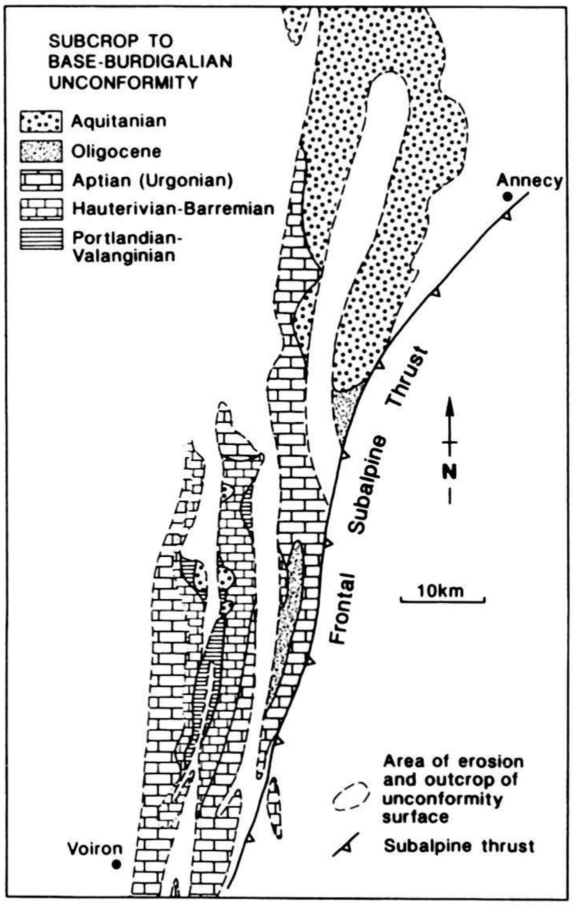


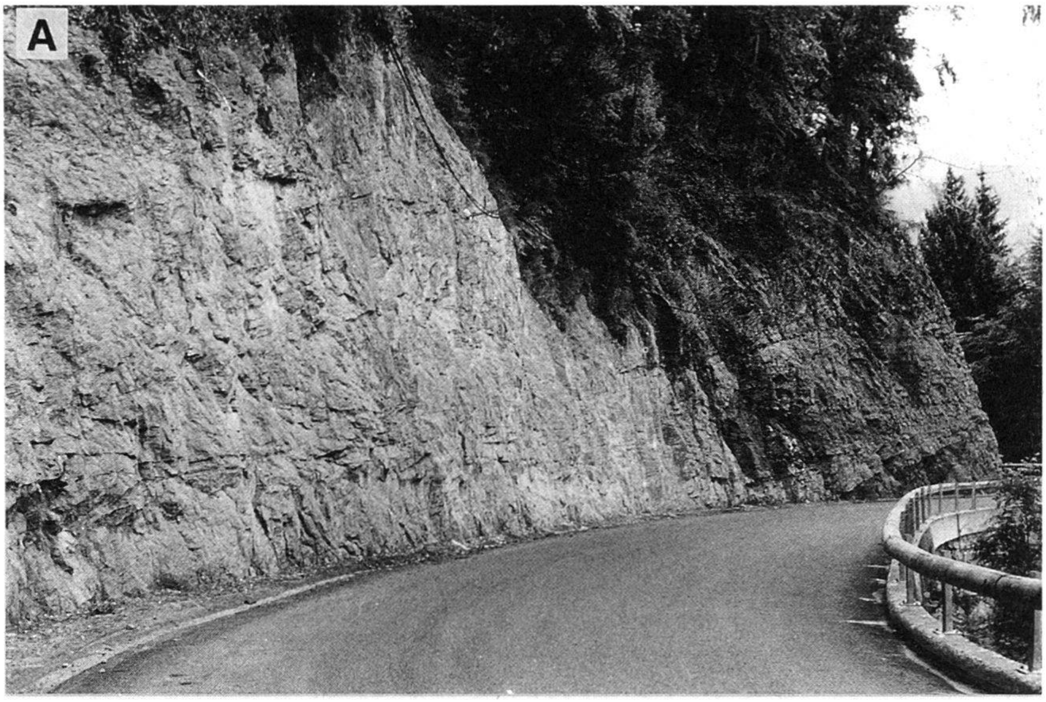
Fig. 7. Subcrop map of the base-Burdigalian unconformity.

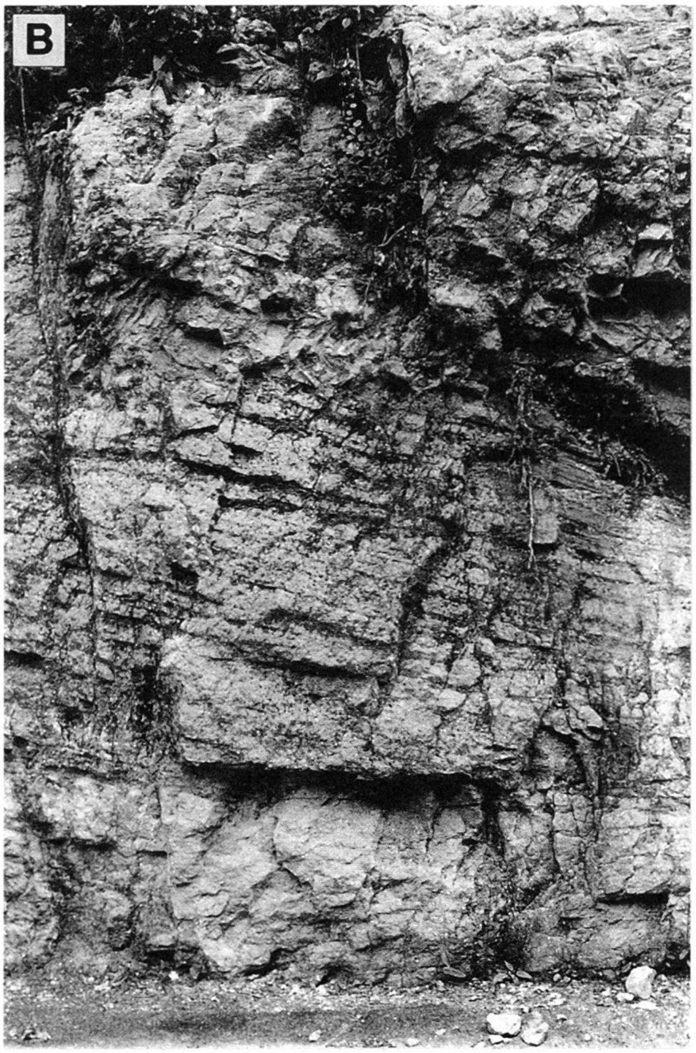
Aquitanian flanking Mont Tournier and Mont L'Epine, showing that prominent topographic highs existed during the late Oligocene and early Miocene (see also Blanc 1991, fig. 73, p 25). The orientation of these paleohighs (approximately N–S) suggests that they were most likely related to faults formed during the period of Rhine-Bresse-Rhône extension (late Eocene–Oligocene), rather than to Pyrenean-provençal compression (early Eocene) which produced E–W folds and faults in the Rhône valley (Gigot & Haccard 1972; Gigot et al. 1974).

4.2 Large-Scale Subtidal Sandwaves of the Basal Transgressive Deposits

The basal deposits, which unconformably overlie the Mesozoic substrate, or with conformable contact on Lower Freshwater Molasse, are giant-scale cross-stratified calcareous sandstones which Meylan (1982) termed *Subtidale Sandwellen* (subtidal sandwaves). These deposits are found at or close to the base of the Tresserve lithosome. They

Fig. 8. Highly asymmetrical subtidal sandwaves exposed in the Chartreuse Massif at 872.0/354.2 on the D45 road about 1.2 km northwest of the village of Corbel. (a) Overview of the 60 m long road section through the large-scale subtidal sandwaves. (b) Large-scale foresets becoming tangential in the toeset region. Height of photograph represents 3.5 to 4 m.





are composed of fine-medium and medium-coarse sandstones with abundant bioclastic material (broken bivalve, echinoid and bryozoan fragments, fish teeth), lignite fragments and glauconite. The glauconite (< 2%) commonly occurs in the form of small amorphous peloids (possibly originally faecal pellets, rare forams, shell debris and mica flakes) which have undergone different stages of evolution (Odin & Matter 1981), and also as a pore-filling cement in shell cavities. Most of the glauconite is thought to be reworked, presumably from an area that had experienced lower sedimentation rates and/or turbulence. Fischer (1987) has shown from samples collected at Génissiat, Fornant and Usses, in the north of the study-area, that most of the glauconite grains are due to transformation from mica. The K–Ar dates of the glauconite pellets are too young (29–35 Ma) to allow for reworking from glauconitic Eocene and Cretaceous (40–90 Ma) parent rocks. Glauconite typically occurs in deposits associated with marine transgression and condensation (Odin 1969; Berg-Madson 1983; Bornhold & Giresse 1985; Donovan et al. 1988; Loutit et al. 1988). The abundance of polycrystalline quartz and the presence of detrital high-pressure index minerals (Mange-Rajetzky & Oberhänsli 1982) indicate an internal Alpine source for much of the sediment.

The large-scale sandwaves can be divided into two types on the basis of internal geometry and bedform morphology:

1. **Highly asymmetrical forms with steep foresets** (Figs. 8, 9, 10): prominent foreset surfaces are long (< 30 m), gently tangential and steeply dipping (25°–30°). Foreset heights are between 4 and 5 m and cross-strata are 5 to 20 cm thick. Occasional thin mud drapes are present on the smooth or rippled toeset surfaces. Packets of foresets are truncated by more gently inclined and sigmoidal bounding surfaces. These are in turn truncated by low-angle bounding surfaces overlain by trough cross-stratified sandstone units. Even higher order surfaces underlie the sandwave complexes and truncate underlying complexes.
2. **Less asymmetrical, composite forms lacking steep, simple foresets** (Fig. 11): the internal structure is composite, with many trough-shaped, gently-inclined, interwoven erosion surfaces bounding bidirectional cross-strata, especially towards the upper half of the preserved stratification sequence. Some master bedding surfaces are gently inclined and dip in one direction.

Palaeocurrent directions taken from foresets within the sandwaves, and from the trough cross-stratified sandstones capping large-scale foreset units, both indicate a migration direction towards the south and SW in the example from the Chartreuse illustrated in Fig. 8. The tight clustering suggests that the sandwave crestlines were not markedly sinuous. Palaeocurrents from all occurrences of the large-scale subtidal sandwaves in the Rhône-Alp region are complex (Fig. 12). Even nearby stations (cf. Figs. 11 and 13) may show markedly different palaeocurrent patterns, with a SE, SW or S net sediment transport in the section south of Seyssel (Fig. 11), and a predominantly N transport at Génissiat, with a subordinate mode to the SW (Fig. 13). Locations between the palaeohighs have longitudinal palaeocurrent directions, either to the NE–E, or S–SW. Locations close to palaeohighs have complex palaeocurrent patterns with significant cross-seaway modes (E–W), suggesting that vigorous tidal currents flowed between islands along the axes of the palaeohighs. These cross-seaway routes may have been older river valleys flooded by the Burdigalian transgression.

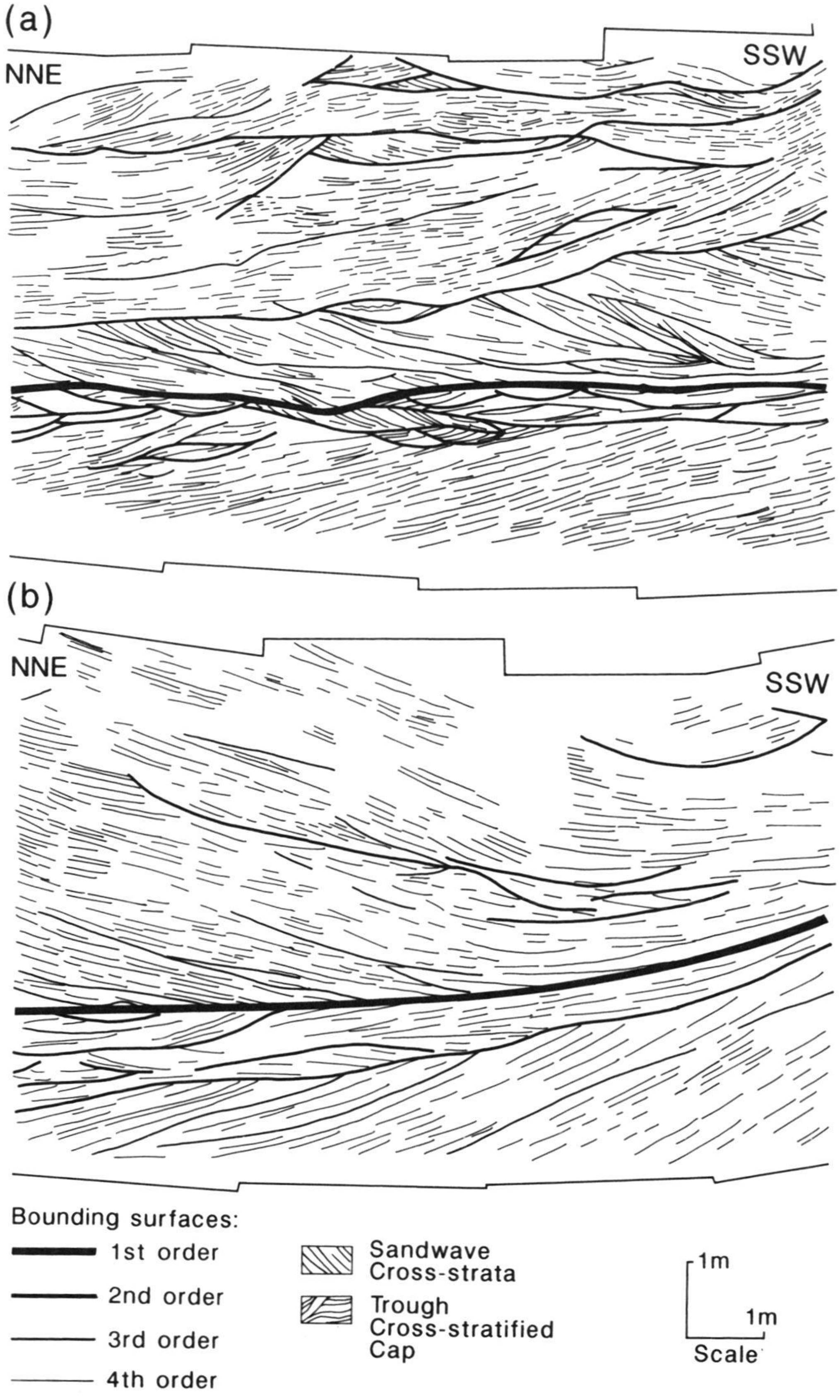


Fig. 9. (a) and (b) Line drawings of parts of the highly asymmetrical sandwave complexes exposed on the roadside close to Corbel (872.0/354.2), showing the hierarchy of bounding surfaces.

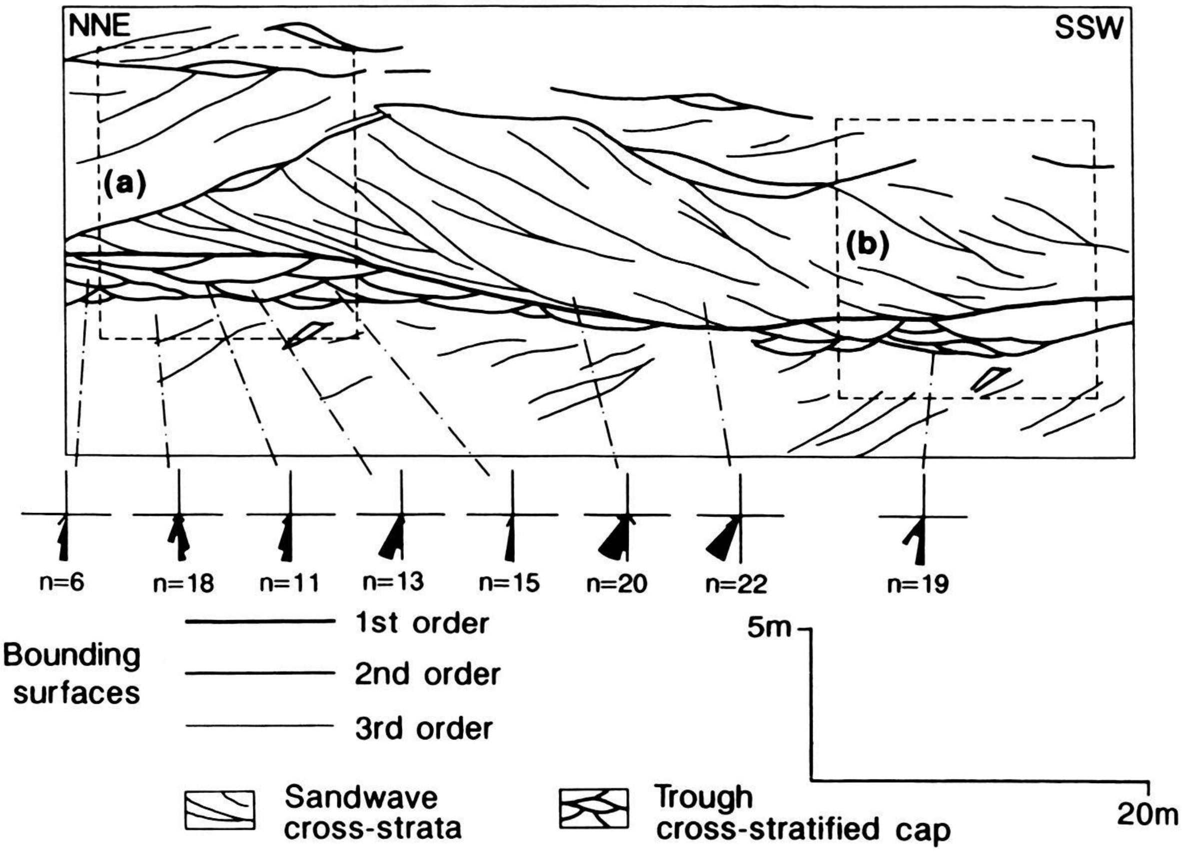


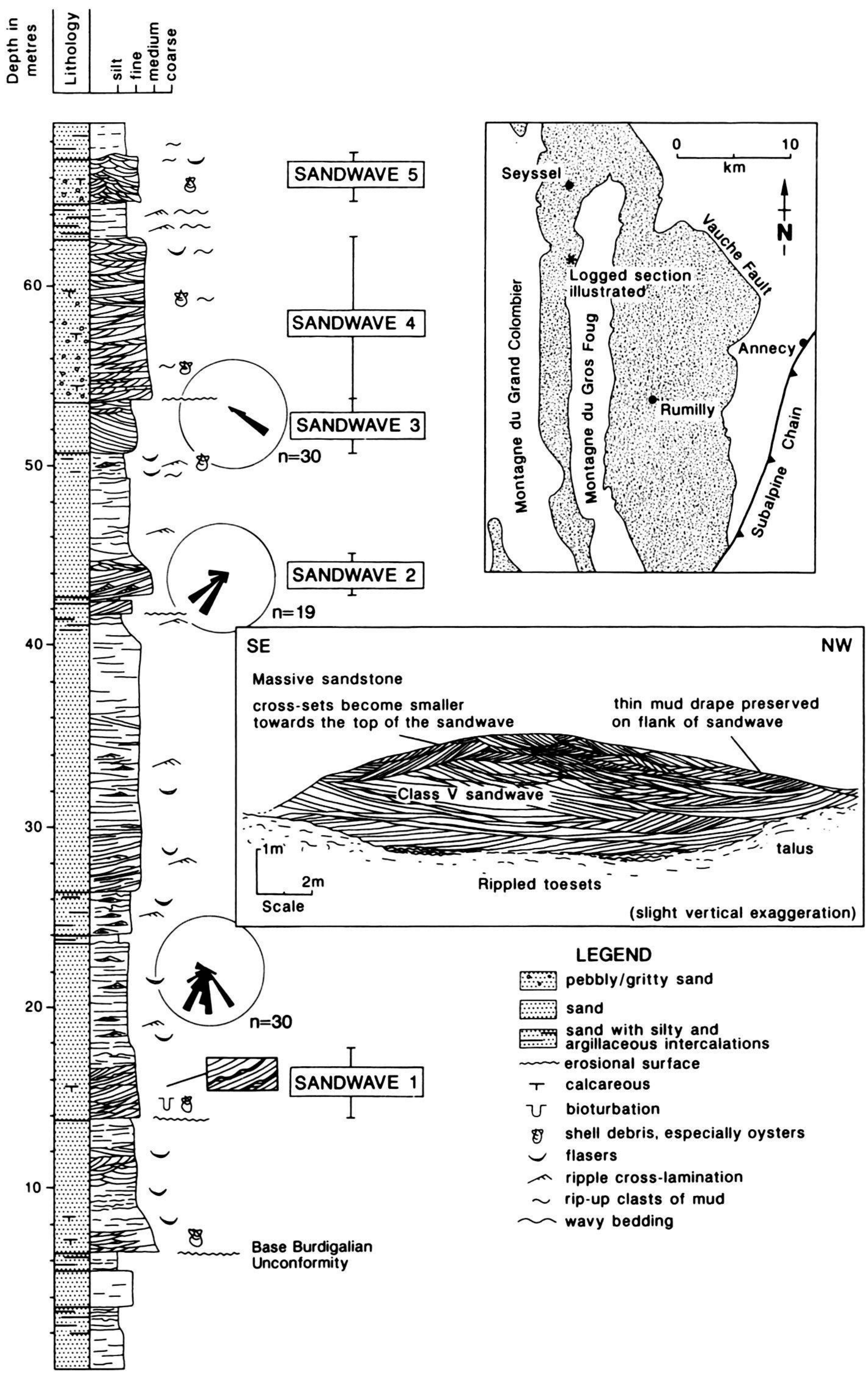
Fig. 10. Simplified diagram of the sandwave complex exposed at the Corbel roadside locality, with palaeocurrent data from foresets and trough cross-stratified cap showing a net sediment transport to the south. Boxes show areas illustrated in Figure 9 (a) and (b). Note vertical exaggeration (approximately x2).

Packets of cross-strata in the asymmetrical sandwaves pass down foreset dip into thinning- and fining-up cycles of fine sandstone and silt couplets, as seen near the top of the Génissiat section (Fig. 13). The basal portions of the cycles are dominated by rippled sandstones with some muddy flasers. The upper portions contain wavily-laminated thinner sandstone beds and continuous muddy drapes. Cycles typically have 15 to 20 couplets.

4.2.1 Interpretation of the Asymmetrical Forms

The unidirectional orientation of the foresets suggests a flow with a strong undirectional net sediment transport. The presence of fine sediment drapes between packets of foresets indicates episodic sand transport, suggesting the operation of tidal currents, each packet of foreset cross-strata most likely being deposited in a neap-spring cycle. Ripple marks are reversely orientated in the toesets (similar features have been described by De Raaf & Boersma 1971; Visser 1980; Allen 1981; Boersma & Terwindt 1981; Home-

Fig. 11. Sedimentary log measured along the stream section 0.5 km east of the village of Ruffieux at 872.2/100.1. The inset contains a line drawing interpretation of the composite sandwave shown as Sandwave 4 in the log.


Depth in metres
Lithology
silt
fine
medium
coarse
60
50
40
30
20
10
SANDWAVE 5
SANDWAVE 4
SANDWAVE 3
SANDWAVE 2
SANDWAVE 1
n=30
n=19
n=30
Base Burdigalian Unconformity
0 10
km
N
Seyssel
Vauche Fault
Logged section illustrated
Montagne du Grand Colombier
Montagne du Gros Foug
Annecy
Rumilly
Subalpine Chain
SE
NW
Massive sandstone
cross-sets become smaller towards the top of the sandwave
thin mud drape preserved on flank of sandwave
Class V sandwave
1m
2m
Scale
Rippled toesets
talus
(slight vertical exaggeration)
LEGEND
pebbly/gritty sand
sand
sand with silty and argillaceous intercalations
erosional surface
calcareous
bioturbation
shell debris, especially oysters
flasers
ripple cross-lamination
rip-up clasts of mud
wavy bedding

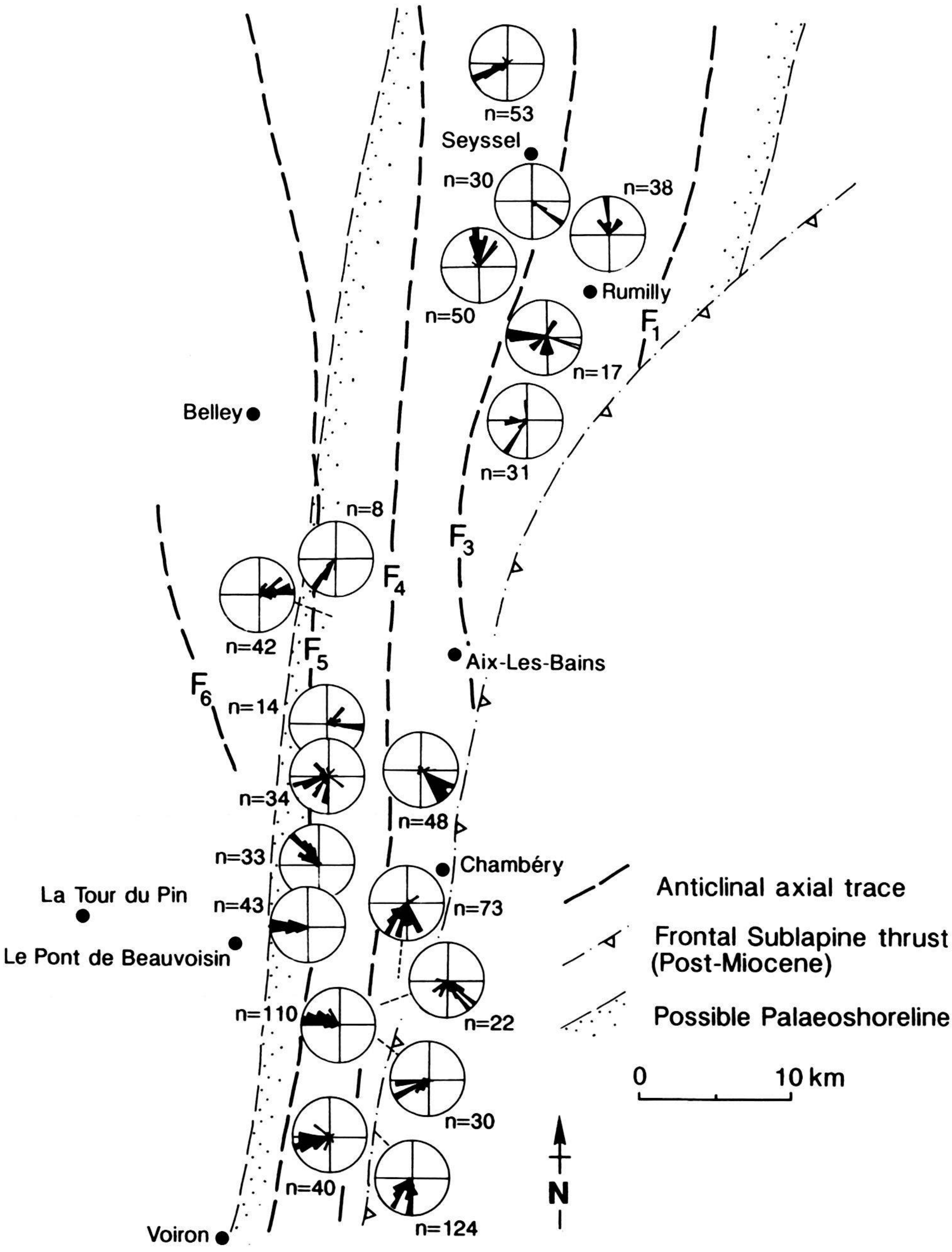


Fig. 12. Palaeocurrent roses of orientations of foreset surfaces in the large-scale subtidal sandwave facies.

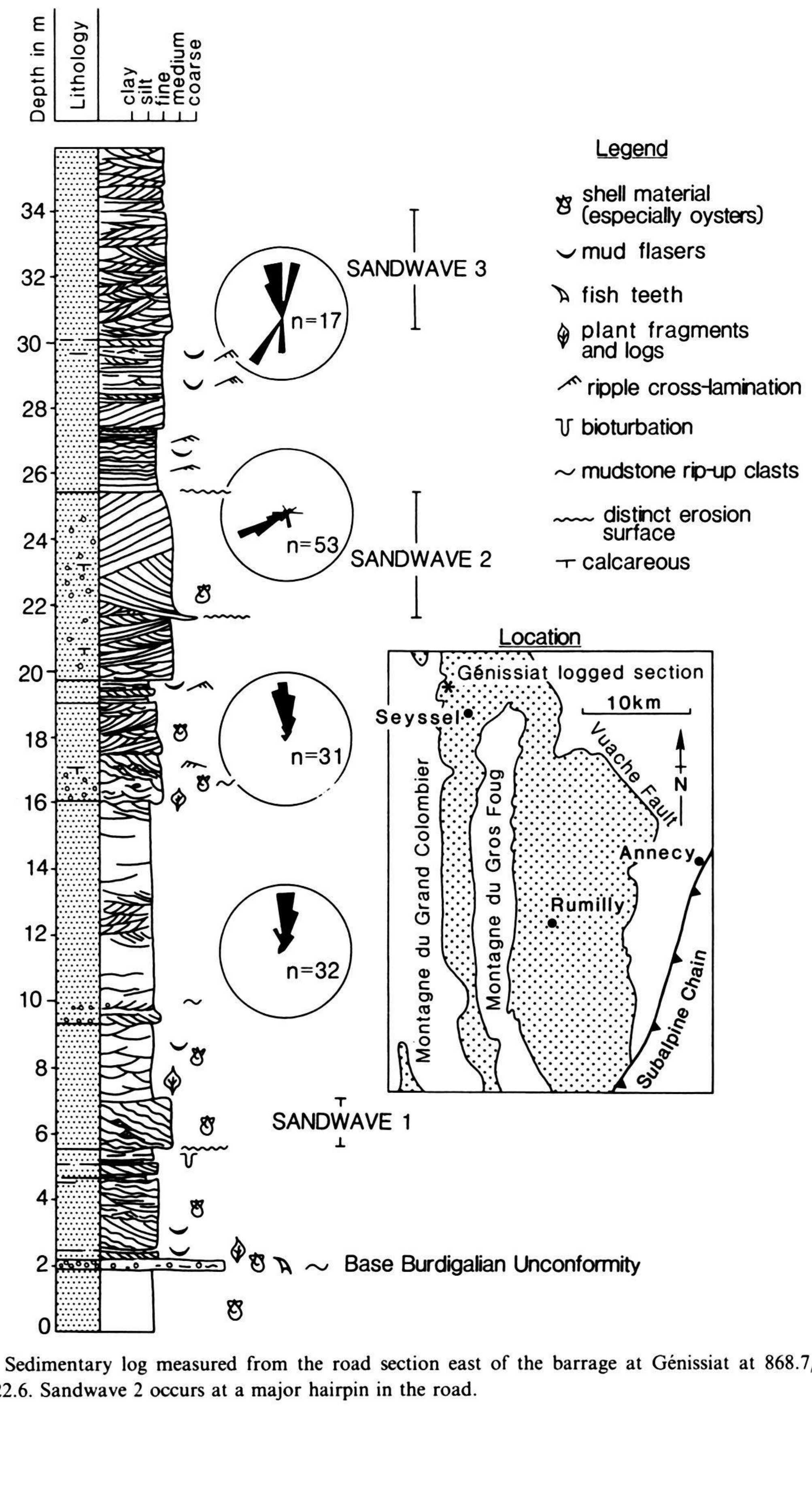


Fig. 13. Sedimentary log measured from the road section east of the barrage at Génissiat at 868.7/121.9 to 869.2/122.6. Sandwave 2 occurs at a major hairpin in the road.

wood & Allen 1981), suggesting reworking by subordinate tidal flows and preservation during neaps. The heights of the preserved foresets suggest relatively deep tidal flows, certainly within the subtidal zone.

The down-dip passage to cyclical heterolithics confirms the tidal control on sediment transport. The couplets appear to represent vertically accreted bundles (*sensu* Kreisa & Moiola 1986), each deposited during a flood-ebb tidal cycle. The thinning- and fining-up nature of the cycles suggests that sediment was preserved on the spring to neap part of the lunar cycle. A strikingly similar association of facies is found in the *Muschel-sandstein* of the OMM of Switzerland described by Allen et al. (1985). The heterolithic facies with a lateral passage to major foresets was clearly deposited in topographic lows between the main positive elements of the subtidal sandwaves.

The less steeply inclined and sigmoidal surfaces truncating packets of foreset cross-strata represent reactivation surfaces created when the sandwave crest was eroded either by stronger reverse currents or storms (Kohsiek & Terwindt 1981) or by a reorientation of the sandwave migration direction.

The trough cross-stratified sandstones overlying the major foreset units *via* a low angle truncation surface represent the migration of small three-dimensional dunes over the stoss-side of the sandwave.

The major erosional surfaces underlying the sandwave complexes, cutting down into underlying trough cross-stratified caps or into the main body of underlying sandwaves, represent the surfaces of climb of the entire sandwave complexes. Such surfaces might be expected to vary from horizontal, where no net sedimentation was taking place, to low angle, where parts or all of the sandwaves were being preserved.

The internal structure and interpreted geometry of the asymmetrical forms are similar to those of subtidal sandwaves studied by McCave (1971) from the southern North Sea, and the large subtidal sandwaves identified on high-resolution seismic reflection profiles of the sea bed of the English Channel (Berné et al. 1988; Berné, Allen et al. 1989) (Fig. 14).

4.2.2 Interpretation of the Composite Forms

The predominance of cross-strata dipping in one orientation, but with the preservation of oppositely orientated structures suggests that the tidal flows were more symmetrical in their sediment transport than in the case of the steep, asymmetrical sandwaves. This is supported by the occurrence of low-angle master bedding surfaces and the lower bedform steepness (Allen 1980a). Clearly, small dunes were able to migrate over the surface of the sandwave both under the dominant tide and the subordinate tide. Flow separation over the rounded sandwave crests was probably limited.

Although the spatial distribution of the two types of sandwave is non-systematic, most of the symmetrical forms are found stratigraphically higher than the strongly asymmetrical forms, which are located close to the basal unconformity.

Large flow-transverse subtidal sandwaves have been described from modern meso-to macro-tidal seas (Lüders 1929, 1936; Van Veen 1935; Stride 1963, 1970; McCave 1971; Bouma et al. 1977; Hine 1977; Allen 1980a, b; Langhorne 1982; Berné et al. 1988). Convincing analogues have been ubiquitously identified from the ancient sedimentary record. The closest analogues, both in terms of structure and stratigraphic position, are

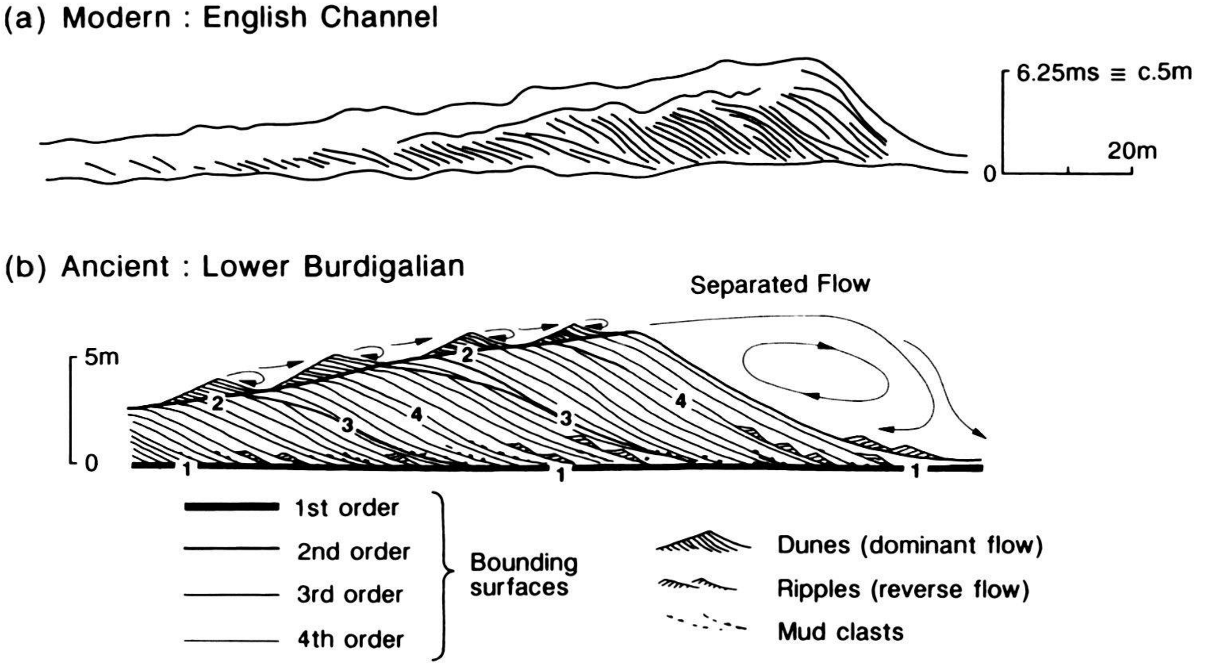


Fig. 14. Comparison, at the same scale, of the subtidal sandwaves observed by shallow high-resolution seismic investigation on the modern sea bed of the English Channel (Berné et al., 1988) and the interpreted bedforms from the large-scale subtidal sandwave facies of the Rhône-Alp region.

the so-called *Muschelsandstein banks* of the Swiss OMM (Allen et al. 1985). These banks occur close to the base-Burdigalian unconformity and are interpreted to have occupied the most offshore facies belt of the seaway.

4.3 Development of a Tide-Dominated Coast (Tresserve and Forezan Lithosomes)

In the east of the study-area, the basal large-scale subtidal sandwaves pass upwards into the sand-dominated part of the Tresserve lithosome, the equivalent of the "*grès glauconieux à ciment calcaire*" of Gidon (1960) and the "*molasse verdâtre et grès calcareux*" of Lamiraux (1977) and Mujito (1981). Macrofauna and microfauna from sandstones in the lithosome (Demarcq 1962; Gigout 1969) suggest a latest Aquitanian to early Burdigalian age (pectinid zone PN3a in Berger 1992). The lithosome is 380 m to 450 m thick in the Chambéry and Rumilly synclines, but thins towards the west, and is absent from sections on the western flank of the Novalaise syncline. The sandstones are very fine to fine grained and are very resistant to weathering owing to the calcareous cement. Towards the top of the lithosome, the calcareous sandstones fine up into bioturbated silty sandstones and siltstones. In addition, it is probable that the lithosome passes westwards (distally from the shoreline) into the bioturbated silts of the Forezan lithosome. The Tresserve lithosome also passes conformably upwards in the Chambéry and Rumilly Synclines into the Forezan lithosome.

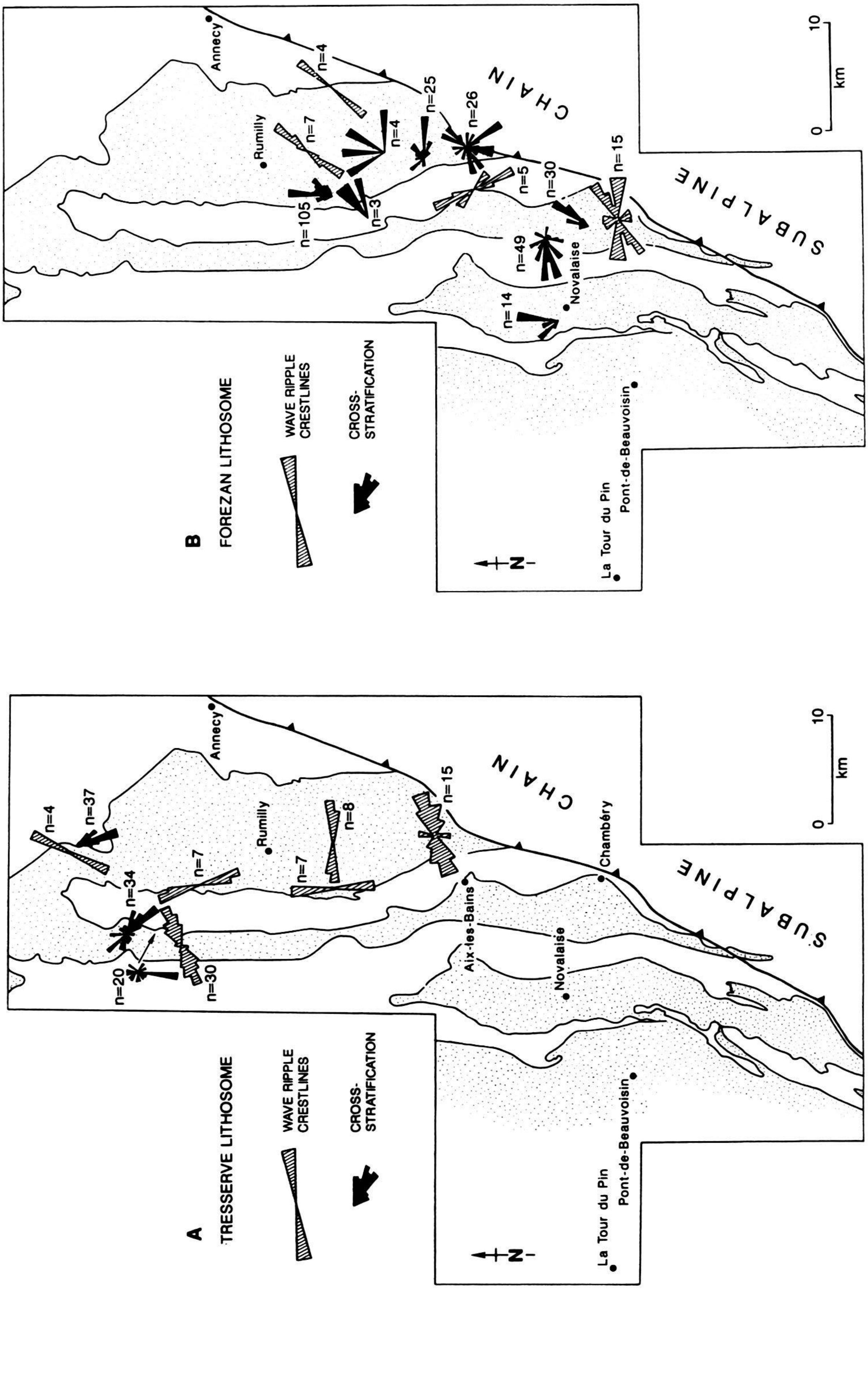
On the western flank of the Novalaise Syncline the Forezan lithosome lies directly upon the basal Burdigalian unconformity, and is thought to be in part age-equivalent of the Tresserve lithosome. The Forezan lithosome passes into a condensed, shelly limestone facies at the western margin of the basin. It is the equivalent of the bioturbated sandy silts of Gidon (1963), Lamiraux (1977), Mujito (1981), Perriaux (1984) and Perriaux et al. (1984), and is at its thickest (230 m) on the eastern flank of the Chambéry Syncline, thinning and fining towards the west (69 m on the eastern margin of the Bas-Dauphiné). It also fines from north to south. The general stratigraphy is of 10–50 m of heavily bioturbated silty sands with some heterolithics at the base, overlain by a number of thick (5–20 m) fining-up units of very fine to fine grained sands to bioturbated silts, passing up into a 5–35 m cap of fine to medium grained sand. The silty sands in particular are rich in lignite, with less common shell fragments and fish teeth. Latreille (1969) identified an abundant microfauna including benthonic and planktonic foraminifera attributed to shallow marine waters deeper than 50 m, whereas the ostracod *Protocytheretta schoeleri*, a key taxa belonging to the early Miocene (pre-Langhian) Biozone A (Carbonnel 1970), lived in open marine waters at depths of up to 30 m. Such microfaunal elements may have been, of course, subject to considerable transport before deposition. The condensed shelly limestones on the western flank of the Novalaise Syncline contain abundant bivalves, echinoderms, bryozoans, barnacles, sponge spicules and foraminifera, suggesting deposition very close to the western shore of the seaway.

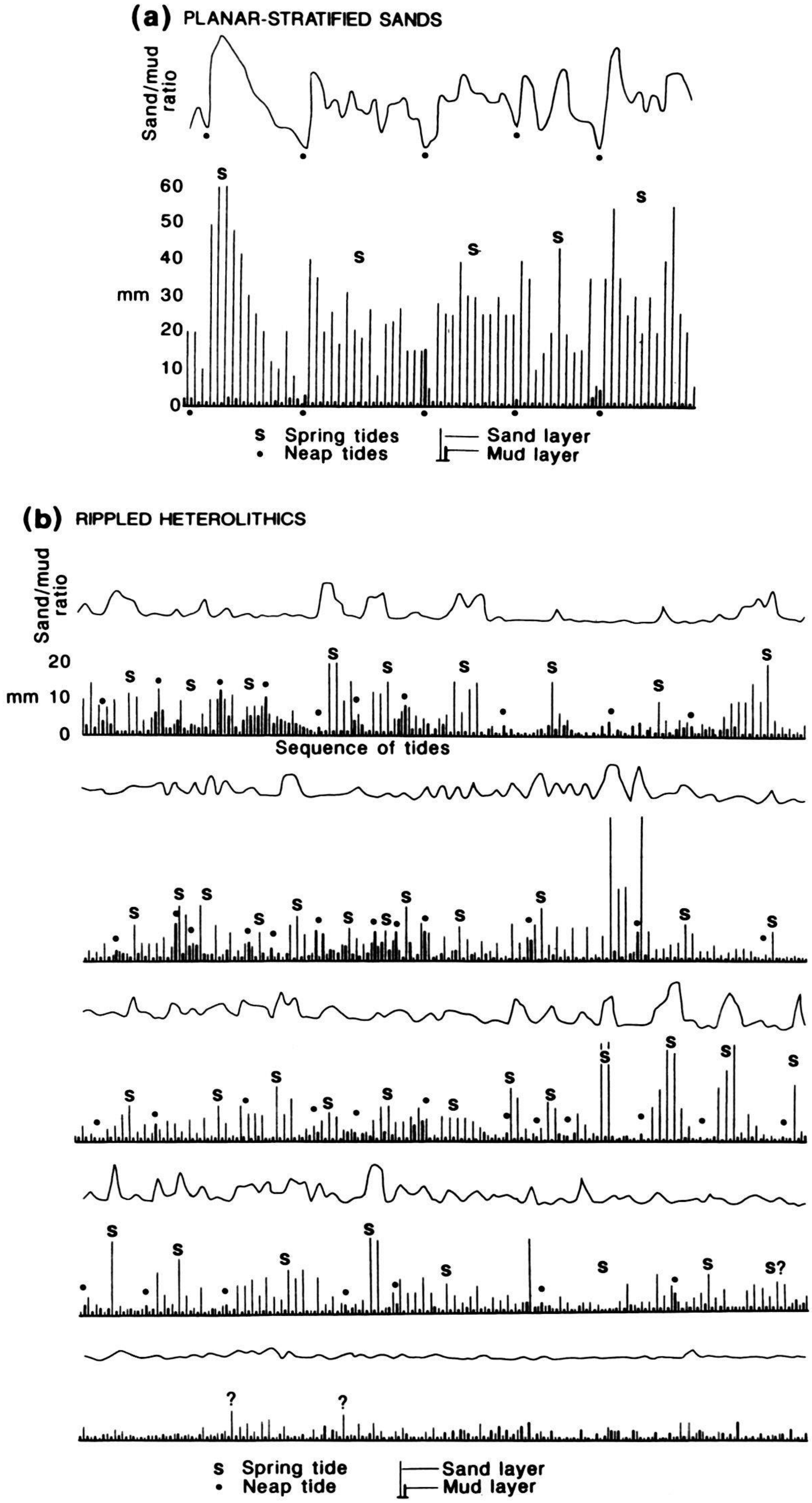
Six marine facies can be identified in the combined Tresserve and Forezan lithosomes, the finer-grained facies being in general more common in the Forezan lithosome, and the sandier facies dominating the Tresserve lithosome:

1. **Cross-Stratified Sandstones** consist of (1) fine to coarse-grained large-scale (4–7 m thick) composite cosets with low-moderate angle (5–15°) bidirectional cross-strata in sets up to 0.4 m thick, bounded by broad trough-shaped erosion surfaces; (2) fine to coarse-grained giant (< 6 m preserved thickness) tabular, unidirectional sets with moderate-high angle (10°–30°) foresets passing laterally into muddy, rippled toesets; and (3) fine to medium grained small- to large-scale trough cross-sets with erosional bounding surfaces lined with thin rippled layers and multiple veneers of silt or mud. The ripples are both symmetrical wave-generated types in the bases of scours and asymmetrical types climbing foresets.
   Subfacies (1) and (2) are similar to the two types of sandwave discussed in section 4.2, and correspond closely to the class IVB and class V sandwaves of Allen (1980a). Subfacies (3) originated under reversing tidal flows generating sinuous-crested dunes. Slack water periods allowed the preservation of mud drapes, particularly in the deep scour pits ahead of dune slip-faces. Subordinate tides locally rippled the bedform lee faces. Background wave activity was able to mould the surfaces of dune scour pits before draping with slack water sediments. The cross-stratified sandstones exhibit a wide spread of palaeocurrent azimuth (Fig. 15), with modes both longitudinally up and down the seaway, and subordinate modes normal to (towards and away from) the inferred palaeocoastline.

2. **Planar-stratified Sandstones** frequently found erosionally truncating the cross-stratified sandstone facies. Decimetre thick tabular beds of very fine to fine grained sandstone are defined by the presence of laterally continuous linsen interbeds. Internally, the tabular beds are composed of planar lamination, with

Fig. 15. Palaeocurrent azimuths from cross-strata (large-scale cross-stratified sands and trough cross-stratified sands) and orientations of wave-ripple crestlines in (A) the Tresserve lithosome and (B) the Forezan lithosome.


B
FOREZAN LITHOSOME
WAVE RIPPLE CRESTLINES
CROSS-STRATIFICATION
N
Annecy
Rumilly
Novalaise
La Tour du Pin
Pont-de-Beauvoisin
CHAIN
SUBALPINE
0 10 km
n=4
n=7
n=4
n=25
n=26
n=105
n=3
n=5
n=30
n=15
n=49
n=14
A
TRESSERVE LITHOSOME
WAVE RIPPLE CRESTLINES
CROSS-STRATIFICATION
N
Annecy
Rumilly
Aix-les-Bains
Chambéry
Novalaise
La Tour du Pin
Pont-de-Beauvoisin
CHAIN
SUBALPINE
0 10 km
n=4
n=37
n=7
n=8
n=7
n=15
n=34
n=20
n=30


(a) PLANAR-STRATIFIED SANDS
Sand/mud ratio
mm
0
10
20
30
40
50
60
s Spring tides
• Neap tides
Sand layer
Mud layer
(b) RIPPLED HETEROLITHICS
Sand/mud ratio
mm
0
10
20
Sequence of tides
s Spring tide
• Neap tide
Sand layer
Mud layer

occasional small scours filled with tangential ripple cross-laminations. Commonly, the bed is made of repeated small-scale cycles (Fig. 16a) comprising parallel-laminated sandstones overlain by rippled layers and then mud drapes.
These small-scale cycles record the passage of dominant tide (parallel-lamination), subordinate tide (ripple cross-lamination) and slackwater fall-out from suspension (drapes). The lack of a drape between the parallel-laminated and ripple cross-laminated divisions may be due to erosion during the rising subordinate tide, to residual turbulence in the water column or, perhaps most likely, to non-deposition in an intertidal environment. The thicker fine sediment drapes in the linsen interbeds represent the weaker flow velocities of neap periods. Similar facies have been described from a modern high-energy tidal sand-flat in the Bay of Fundy (Dalrymple et al. 1990), though drapes are absent because of the wave activity at high tide. Similar sequences have been described from tidal flat deposits in the Mississippian of Illinois (Wescott 1982) and the Cretaceous Curtis Formation of Utah (Kreisa & Moiola 1986).

3. **Rippled Heterolithics** consisting of sequences of interbedded mm to cm thick cross-laminated fine sandstones with wave rippled top surfaces and mm-scale silts in thinning- and fining-up packets.
Individual sand-silt couplets represent one tidal (ebb-flood) cycle and therefore may be termed bundles, the thinning- and fining-up of the packets being due to variations in sediment transport through neap-spring cycles (Fig. 16b). Similar bundles and cycles in heterolithics were described by Tessier & Gigot (1989) and Tessier et al. (1989) from the Burdigalian of the Digne area, and Assemat (1991) from the Miocene of the Rumilly Syncline. Although an intertidal origin is possible, the lateral transition from rippled heterolithics into sandwave toesets make it more likely that this facies was deposited between sand shoals in subtidal depths or in subtidal channels. Wave ripple crestlines are mostly E–W, suggesting waves travelling longitudinally in the seaway unaffected by coastal refraction.

4. **Bioturbated Silty Sandstones and Massive Sandstones** are found principally towards the base of the Forezan lithosome. They lack sedimentary structures, consisting of a poorly sorted mixture of dark grey silt to fine-medium sand with dispersed lignite flakes, or massive very fine to fine sands mottled by lignite flakes. Cylindrical traces of *Thalassinoides* are locally visible, but the bioturbation is in general homogeneous and pervasive. An environment of relatively low sediment delivery rate in the subtidal zone below fairweather wave-base is inferred.

5. **Laminated Siltstones** contain millimetre-thick very fine sand-starved ripple lenses (Kaneko & Honji 1979) and pinstripes. Siltstone laminae have sinuous feeding traces of *Didymaulichnus* (Hakes 1976) and *Aulichnites*. A quiet environment starved of coarse sediment supply is indicated.

Fig. 16. Measured bundle sequences from the Tresserve lithosome. (a) Planar stratified sands facies at 873.1/119.0 on the D31 road 5 km SW of Mons. (b) Rippled heterolithic facies from stream section at 875.4/096.0, 2 km NE of Cessens.

6. **Intraclast Conglomerate Sandstones** occur in channel-form bodies up to 3 m thick. Convoluted fine-grained intraclasts up to pebble-boulder size occur as a chaotic mix in a sandy matrix. The intraclasts are thought to have been derived locally by bank collapse of probably intertidal channels.

A coastal tide-dominated environment is indicated by the association of facies. The subtidal shoals with intertidal crests either existed in estuarine embayments or were detached from the coastline. Since most palaeocurrents from the large-scale cross-stratified sands (subtidal sandwaves) and trough cross-stratified sands (tidal dunes) are oriented along the seaway, a detached origin is preferred. The laminated silts are associated with nearshore and intertidal facies, as well as with interpreted intertidal channel-fills. They therefore may represent lagoonal deposits protected by a barrier, or estuarine mudflat deposits.

The influence of palaeohighs on marine sedimentation decreased through time. In the Tresserve lithosome limestone clasts are found close to the positions of palaeohighs, but in the Forezan lithosome this association is not found. The islands that had characterized the early Burdigalian phase of the seaway had probably been submerged by the late Burdigalian.

Within the Forezan, the facies are considerably sandier in the northern half of the study area (Rumilly Syncline). A fluvial feeder system may have entered the seaway in this region. Tardy & Doudoux (1984) and Huggenberger & Wildi (1991) have drawn attention to an important fault zone within the external part of the Bornes Massif (10 km east of Annecy), which may have controlled dispersal from the Alpine hinterland to the seaway. To the south, the coast may have been lower gradient, with wider intertidal zones. Bioturbation is also more common in the south, suggesting lower sedimentation rates away from the main sediment delivery system in the north.

4.4 An Interlude of Low Energy Conditions (Montaugier Lithosome)

The Montaugier lithosome, which conformably overlies the Forezan lithosome in the Chambéry Syncline, Novalaise Syncline and Rumilly Syncline, marks a basin-wide change to muddy sedimentation. It is not found in the Bas-Dauphiné. The Montaugier lithosome was first described by Demarcq (1962), and also by Lamiraux (1977) who named it "*Les marnes bleues argileuses*". It thins from east (48 m – 72 m in the Chambéry Syncline) and north (25 m – 65 m in the Rumilly Syncline) to west (9 m – 30 m in the Novalaise Syncline). The base of the lithosome is middle to late Burdigalian in age, based on the occurrence of calcareous nannoplankton (NN3 – NN4 zones) (Lamiraux 1977). The top is more problematical. If the Montaugier lithosome is correlated with the "*Marnes bleues de Faucon*" (Demarcq 1962), 100 km to the south, which range up to the early Langhian, and the unit is not strongly diachronous, the age range of the Montaugier lithosome is mid-late Burdigalian to early Langhian. The sandy molasse in the vicinity of Voreppe (close to Grenoble) may be a time-equivalent, suggesting that the sea bed coarsened to the south, towards a sediment effluent entering the basin from a conduit close to the present Isère valley. However, exact synchroneity cannot be demonstrated.

The lithology is of thick units (tens of metres) of bioturbated clayey marl, and rare, thinner units (< few m) of bioturbated silty sandstone. The lithosome coarsens up in its

Fig. 17. Laminated blue-grey clayey-marl with millimetre thick laminations of very fine sand from the Montaugier lithosome. Lens cap is 50 mm.

upper few metres, where very thin fine grained sandstone beds occur. Most of the assemblage of foraminifera lived in water depths of 0–50 m, although a number of deeper water forms were probably swept in from the Tethys Ocean to the south (Berger 1985). There are three main facies:

1. **Laminated Blue-Grey Clayey-Marl** containing rare, very thin laminae and starved linsen of very fine sand, locally bioturbated (Fig. 17). Low energy background conditions allowed the fall-out from suspension of clay grade material. Intermittent sand supply may have been related to storm winnowing of nearshore areas and transport offshore. However, detailed measurement of sandstone and mudstone layer thicknesses (Fig. 18) show that there may also be a tidal signature, albeit diffuse, in the sediment supply. Sand layers rapidly decrease in abundance above the base of the lithosome, suggesting deepening of water or raising of storm wave base through time.

2. **Stratified Sandstones,** found only in the Rumilly Syncline, composed of thick beds (< 10 m) of bioturbated dark grey silty, very fine sandstone containing thin planar and ripple cross-laminations and muddy flasers. Closer proximity to a source of silt and sand and an environment capable of winnowing out the clay grade material is indicated in the north of the study area.

3. **Interbedded Fine Sandstones and Siltstones** occur in the top 10 m of the lithosome in a coarsening-up succession. The sandstone beds (varying from 10 cm at the base

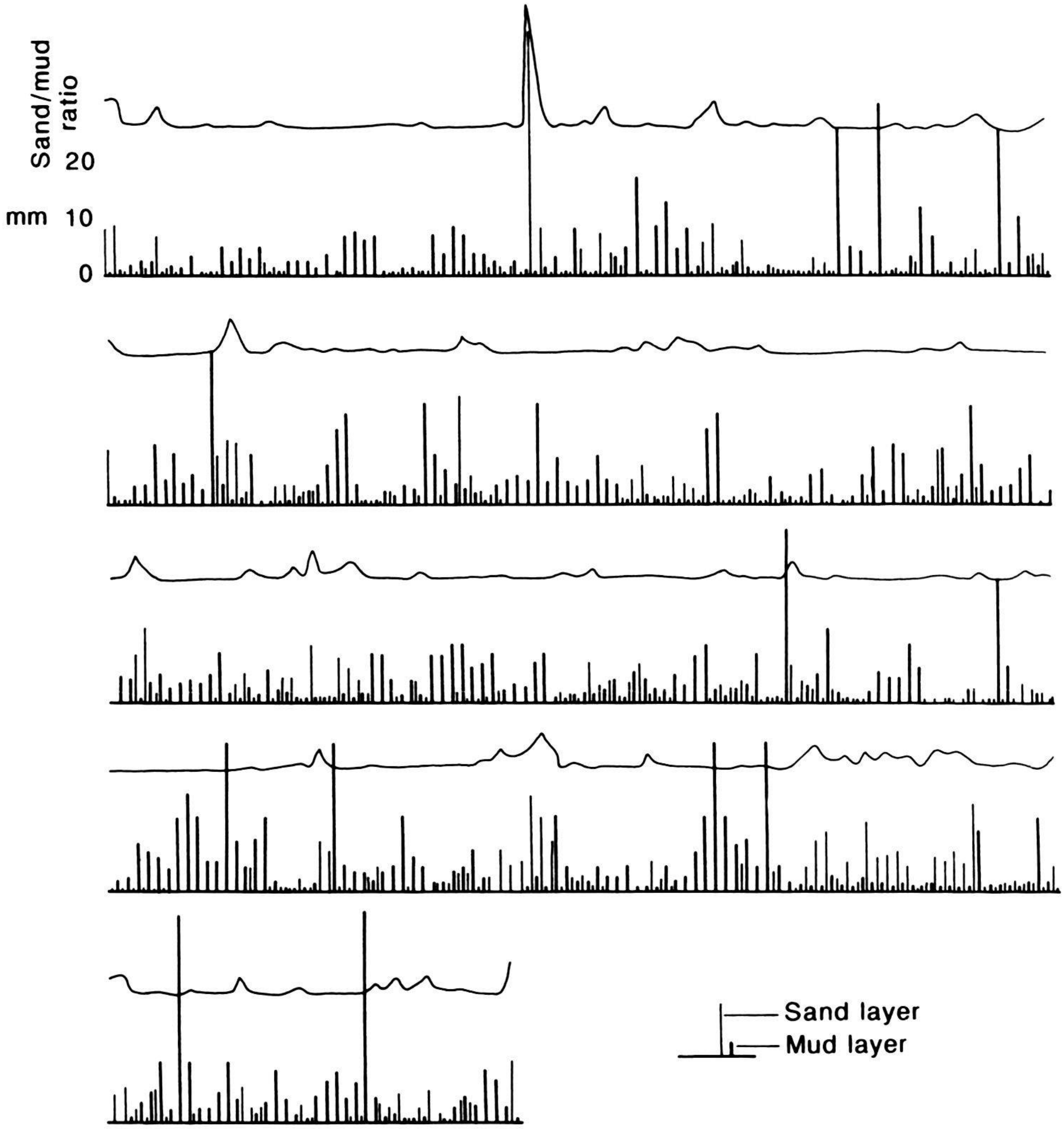


Fig. 18. Thickness of alternating sand and mud from the laminated blue-grey clayey-marl facies of the Montaugier lithosome. Exceptional peaks in the sand/mud ratio are probably due to storm input of sand to the depositional site. Mud bed thickness trends show good cyclicity in some parts of the measured section, attributed to neap-spring variations in sediment transport rates, whereas other parts of the section have little recognizable cyclicity.

to 20 cm thick at the top) have planar bases and contain ripple cross-lamination. Interbedded siltstones contain linsen and pinstripes of very fine sandstone.

The marked change in environments of deposition between the tide-dominated coastal environments of the Forezan lithosome and the quiet waters, probably below storm wave base, of the Montaugier lithosome can be interpreted in two ways. Firstly, rapid deepening of water may have taken the sea bed below storm wave base. Secondly, a narrowing of basin width may have raised storm wave base by its reduction of fetch and therefore wave period. The Montaugier lithosome does indeed pinch out along the western flank of the Novalaise Syncline approximately 10 km east of the pinch out of the underlying lithosome, indicating basin narrowing. An eastward (basinward) shift in the coastline during a relative sea level fall does not appear to be compatible with deepening of sea level in the basin. The observed features cannot be explained solely by a relative sea level rise (see section 5.3.2).

Although no coarse clastic sediments are found in the Montaugier lithosome, the presence of stratified sandstones in the Rumilly area in the north-east, the presence of sandstone-dominated sections in the Villard-de-Lans and Monta Synclines on the eastern side of the seaway reported by Latreille (1969), and the correlation with the Voreppe sandstones to the south-east in the southern Chartreuse (Demarcq 1962; Bocquet 1966; Nicolet 1979; Perriaux 1984) suggest that the Montaugier sea bed coarsened towards the Alpine coast. There are a number of partly analogous present-day and Pleistocene muddy shelf depositional systems, such as the East China Sea (Nittrouer et al. 1984), and Washington-Oregon-northern Californian shelf (Nittrouer et al. 1979; Nittrouer & Sternberg 1981; Leithold 1989). A closer modern analogue may be the narrow tidal sea of the Skagerrak, a strait linking Baltic Sea waters with the North Sea (Eisma 1981). Fine-grained sediment transported into the strait is derived from the rivers Thames, Meuse, Weser and Elbe. Mud deposits are found on the northern side of the strait where water depths (< 100 m) are greatest, sandy sediments being found along the shallower southern side (Van Weering 1981; Van Weering et al. 1973). In the peri-Alpine seaway in eastern France, sediment entry points may have been along a valley close to Voiron-Grenoble in the south, and close to Annecy in the north.

4.5 Return to a Tide-Dominated Coast in a High-Energy Sea (Grésy and Pont-de-Beauvoisin Lithosomes)

The Grésy lithosome: this conformably overlies the muddy deposits of the Montaugier lithosome in the Chambéry, Novalaise and Rumilly Synclines, but, further to the west, it first disconformably overlies the Forezan lithosome with the Montaugier missing, and then lies directly on Oligo-Miocene continental sediments in boreholes in the Bas-Dauphiné. This time period is therefore one of marked expansion of the basin towards the west. The lithosome is equivalent to the "*grès verts*" (greensands) and the overlying "*ensemble marnosilto-gréseux* à Ostrea crassissima" (marly-silty-sandy group containing *Ostrea crassissima*) described by Demarcq (1962) and Lamiraux (1977) in the Chambéry and Novalaise Synclines, but such a subdivision is very difficult to distinguish in the field. Lamiraux (1977) believed the top of the lithosome to be gradational upwards and laterally into the overlying sandstone and gravel-dominated lithosome (Pont-de-Beauvoisin).

The Grésy lithosome thickens from the east (250 m – 310 m in the Chambéry Syncline; 350 m in the Rumilly Syncline) to the west (670 m in the Novalaise Syncline), but then thins rapidly onto the eastern flank of the Bas-Dauphiné where it measures just 56 m. A second, subordinate depocentre is found further west in the Bas-Dauphiné, since boreholes from La Tour-du-Pin (F1, Fig. 2) and 14 km to the south have sections of Grésy lithosome 100 m and 150 m thick respectively. The precise age is once again difficult to ascertain because of poor biostratigraphy. The bivalve *Ostrea crassissima* is a common faunal element in the Swiss "Helvetien" where it is pre-Langhian (NN3 to possibly NN4). The lithosome is believed to be equivalent to the "*sables granitiques de Solaise*" of the *Lyonnaise* region and of marine sands as far north as Bourg in the southern Bresse graben (Demarcq 1962; Latreille 1969).

The internal structure of the Grésy lithosome is of units of very fine to medium grained sand (5–50 m thick) with sharp, erosional lower boundaries, interspersed with thinner units of interlaminated very fine sand and silt. A single 26 m-thick pebbly conglomerate (Cognin Member) is found in the Chambéry Syncline.

The Pont-de-Beauvoisin lithosome: this is only well exposed in the south of the study area within the Chambéry Syncline, the Novalaise Syncline and the western flank of the Bas-Dauphiné. Lory (1860) and several works by Douxami (1895–99) originally described the sandy sequence around the town of Le Pont-de-Beauvoisin. A Serravallian age was attributed to these "*Sables de Pont de Beauvoisin*" by Demarcq (1962) and Latreille (1969). The first occurrences of *Orbulina* in the Pont-de-Beauvoisin sandstones (N8 planktic foraminifera zone or younger), implies that the lithosome is Langhian and perhaps Serravallian. Mujito (1981) concluded that the "*Conglomérats de Chamoux*" found towards the east were time-equivalents of the Pont-de-Beauvoisin Sandstones. Lamiraux (1977) integrated lithostratigraphic correlation, petrography and biostratigraphy of the lithosome from the Chambéry Syncline to the Bas-Dauphiné. The lower component of the lithosome, the Pont-de-Beauvoisin Sandstones becomes more pebbly upwards and thickens westwards. The upper component, the Chamoux Conglomerate, thins westwards from the Chambéry Syncline and does not occur further west than the Novalaise Syncline (Fig. 4a).

In the Chambéry Syncline, the Pont-de-Beauvoisin lithosome consists of 150 m of thickly-bedded Chamoux Conglomerate, which rapidly coarsens up from a 20 m-thick basal coarse sandstone unit. In the Novalaise Syncline, however, the basal sandy facies (Pont-de-Beauvoisin Sandstones or "*Sables jaunes*" of Lamiraux 1977) is at least 90 m thick, coarsens towards its top, and is then abruptly overlain by Chamoux Conglomerate ("*Conglomérats de la Genaz*" of Lamiraux 1977). Further west in the Bas-Dauphiné, the Pont-de-Beauvoisin Sandstones ("*Sables de Pont-de-Beauvoisin*" of Lamiraux 1977) make up the whole of the lithosome, and even further west in the region of La Tour-du-Pin, the lithosome measures 350 m and 400 m in thickness in boreholes (F1, Fig. 2). The lithosome therefore fines from the Alpine front in the east towards the depocentre in the west. The Chamoux Conglomerate is a coarse-grained wedge that lenses out before the eastern edge of the Bas-Dauphiné.

The Pont-de-Beauvoisin lithosome is thought to be time-equivalent of the "*Sables et grès stériles de Chabrillan*" (200 m) in the Crest Basin, the "*Conglomérats de Voreppe*" in the southern Chartreuse and Vercors, the "*Sables de Montchenu*" (60 m) in the southern Bas-Dauphiné, and the "*Sables de Saint-Fons*" (150 m) in the Lyonnaise region (Demarcq 1962; Bocquet 1966; Latreille 1969; Mortaz 1977; Lamiraux 1977; Nicolet 1979). It also appears to be equivalent to the "Yellow Molasse" (MJ) of the Digne Basin (Crumeyrolle et al. 1991). Bivalves (*Natica helicina, Chlamys gentoni, Patella sp.*), terrestrial gastropods (*Planorbis sp., Limnea sp.*) (Demarcq 1962), benthic and planktic normal-salinity foraminifera (*Cibicides lobatulus, Globigerinoides trilobus*), and the brackish form *Ammonia becarii* (Latreille 1969) suggest mixing of the assemblage from different sources (Berger 1985) into a nearshore region.

The Pont-de-Beauvoisin lithosome is overall coarser-grained than the Grésy lithosome; the Pont-de-Beauvoisin Sandstones are composed of fine to medium grained sands, locally in 10–15 m-thick fining-up cycles.

A large number of sedimentary facies can be differentiated (Bass 1991) from the two lithosomes; these can be grouped primarily into sandy facies and heterolithic facies, summarized below.

4.5.1 Sandy Facies

Five sandy facies can be recognized, based primarily on the geometry of the cross-stratification:

1. **Large-Scale Cross-Stratified Sandstones** in fine to medium grained units up to 5 m thick, composed of planar tabular cross-stratification in sets up to 1.5 m thick (Fig. 19). Mud clasts and chips, lignite flakes, shell debris and rare fish teeth occur. Low to medium angle (10°–20°) planar and tangential foresets pass into toesets with reverse-orientated ripples and overlying muddy drapes, and in some cases pass laterally into interbedded sandstone and siltstone heterolithic facies. Systematic variations in foreset angle and shape associated with shallow erosional reactivation surfaces or pause-planes (Fig. 20) (Boersma & Terwindt 1981) are found. The episodic migration of the bedforms, either punctuated by reverse-flow ripple migration and fall-out of fines, or the cutting of reactivation surfaces, demonstrates a tidal control on sediment transport in small sandwaves. Such sandwaves probably occupied subtidal channels. The systematic variations in foreset slope reflect changes in the rate of suspended load fall-out onto the bed through time (Kohsiek & Terwindt 1981), perhaps related to flow deceleration through a tidal cycle. The palaeocurrent azimuths show no clear preferred orientation, although shore-parallel currents appear to have dominated in some cases (Fig. 21 A). Some palaeocurrent modes are to the SE to E and the W, suggesting the influence of coastal tidal embayments and inlets on bedform migration directions.

2. **Trough Cross-Stratified Sandstones** in very fine to fine (occasionally medium) grained, laterally extensive (several tens of metres) units from 1 m to 30 m thick, with abundant mud rip-up clasts, especially along their erosive bases (Fig. 19). The cross-stratification is medium to high angle and tangential, and foresets are commonly covered by reverse-orientated climbing asymmetrical ripple cross-lamination capped by a single mud drape. Foresets pass laterally into distinct highly scoured troughs (1 m–5 m width, 0.2 m depth) which have a fill of numerous sand/silt couplets, occasionally with symmetrical wave ripple-marks, lensing out towards the edge of the trough. Exceptionally preserved form-sets show that the dunes were 0.4 to 0.5 m in height and 3 to 4 m in wavelength (Fig. 22).
   The cross-stratification was produced by the migration of sinuous-crested dunes under the dominant tide, followed by ripple migration under the subordinate tide and slackwater deposition. Scour pits ahead of bedform slipfaces accumulated fine, rippled sediments and drapes, probably especially during neaps when sedi-

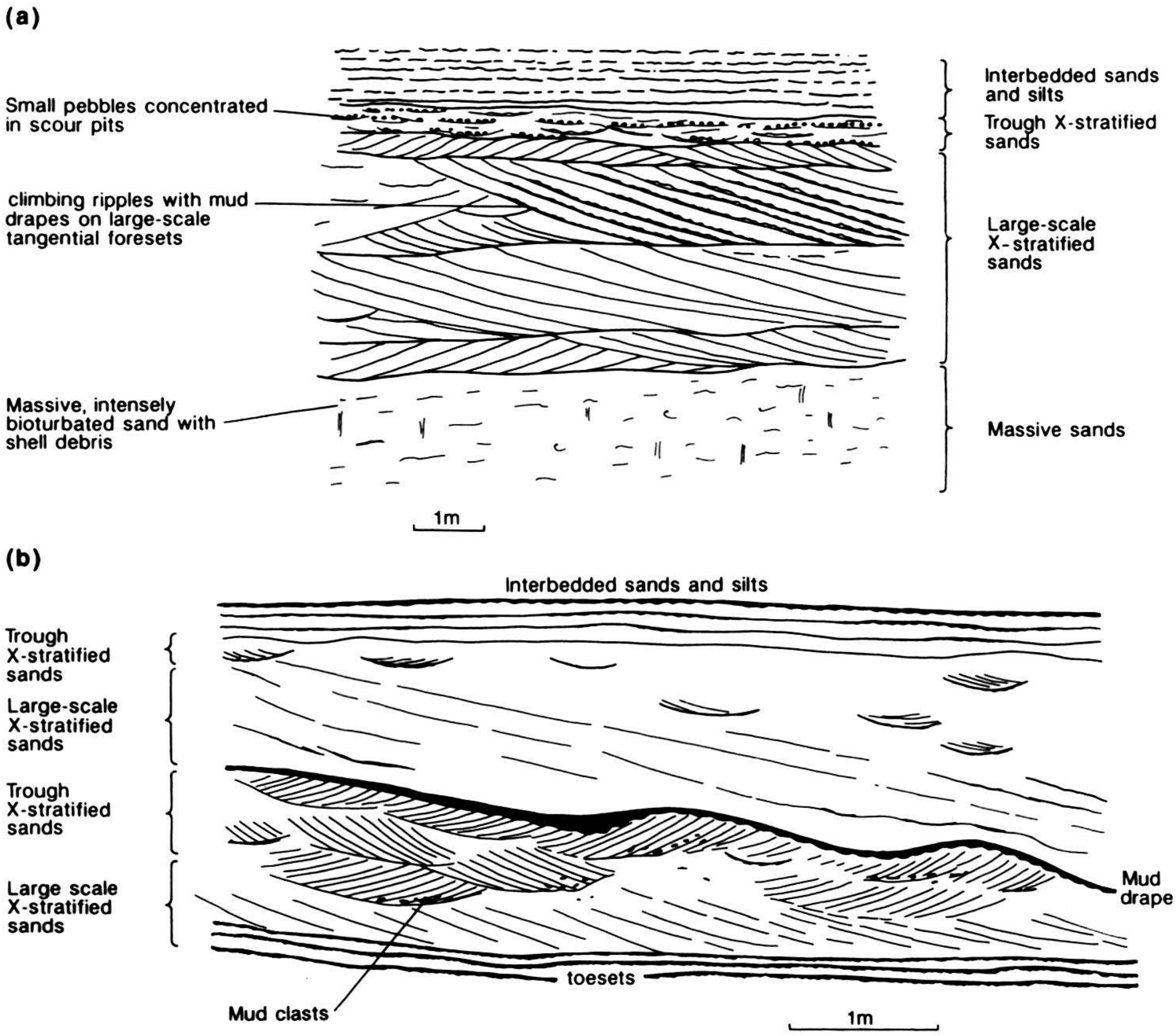


Fig. 19. Facies sequences from the Grésy lithosome. (a) Intertidal mixed-flat massive (bioturbated) sands are erosively overlain by the large-scale cross-stratification of a subtidal channel-fill and finally interbedded sands and silts of the deeper subtidal zone. From "le Sierre" stream section at 882.6/088.5, 3 km ENE of Grésy-sur-Aix. (b) Tidal sandwaves with superimposed dunes sandwiched between interbedded sands and silts representing the toesets of subtidal shoals or sandwave complexes. From "les Merdaret" stream section at 867.0/077.8 close to the D921 road. From field sketches with same vertical and horizontal scales.

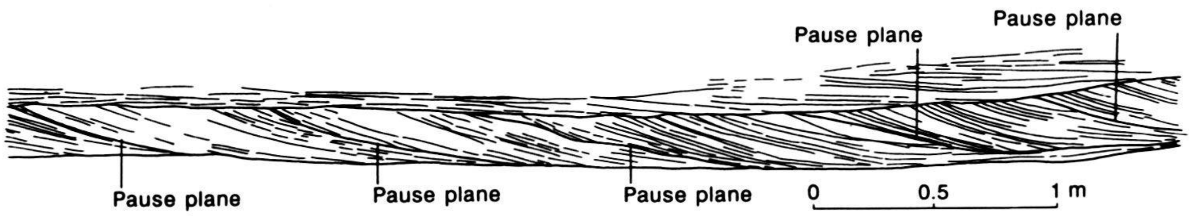
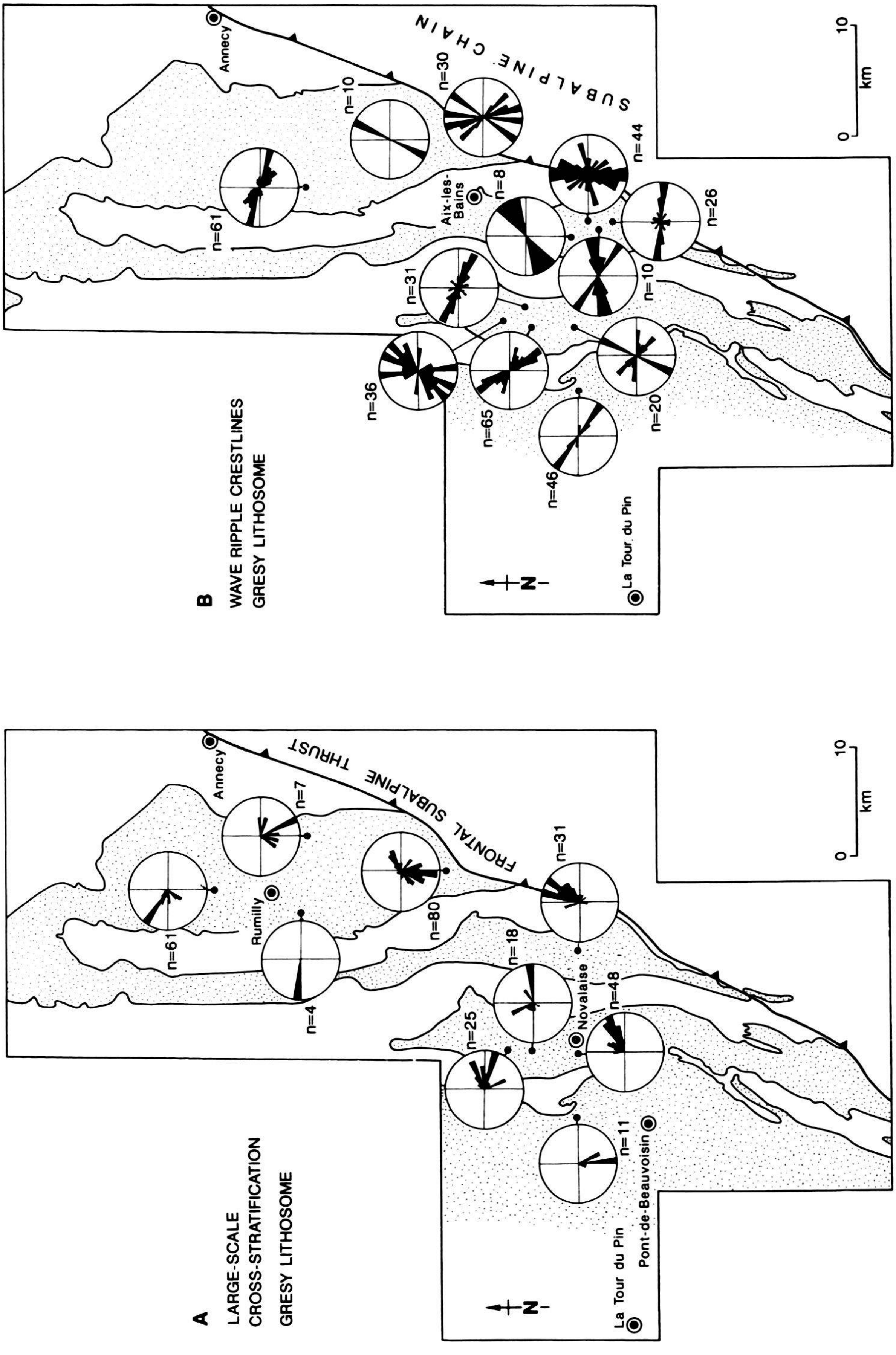


Fig. 20. Line drawing of a tabular set deposited by a small straight crested sandwave from the Pont de Beauvoisin lithosome. In the field of view of the diagram the sandwave experienced sediment transport in four phases, separated by pauses in frontal accretion. From 860.7/066.5 beside the D916a road 1.5 km N of Le Pont-de-Beauvoisin.

Fig. 21. Palaeocurrents from the Grésy lithosome. (A) Large-scale and trough cross-stratification azimuths. (B) Orientation of wave-ripple crestlines.


B
WAVE RIPPLE CRESTLINES
GRESY LITHOSOME
SUBALPINE CHAIN
Annecy
Aix-les-Bains
La Tour du Pin
N
n=61
n=10
n=30
n=44
n=8
n=26
n=31
n=10
n=36
n=65
n=20
n=46
0
km
10
A
LARGE-SCALE
CROSS-STRATIFICATION
GRESY LITHOSOME
FRONTAL SUBALPINE THRUST
Annecy
Rumilly
Novalaise
Pont-de-Beauvoisin
La Tour du Pin
N
n=61
n=7
n=80
n=4
n=31
n=18
n=48
n=25
n=11
0
km
10

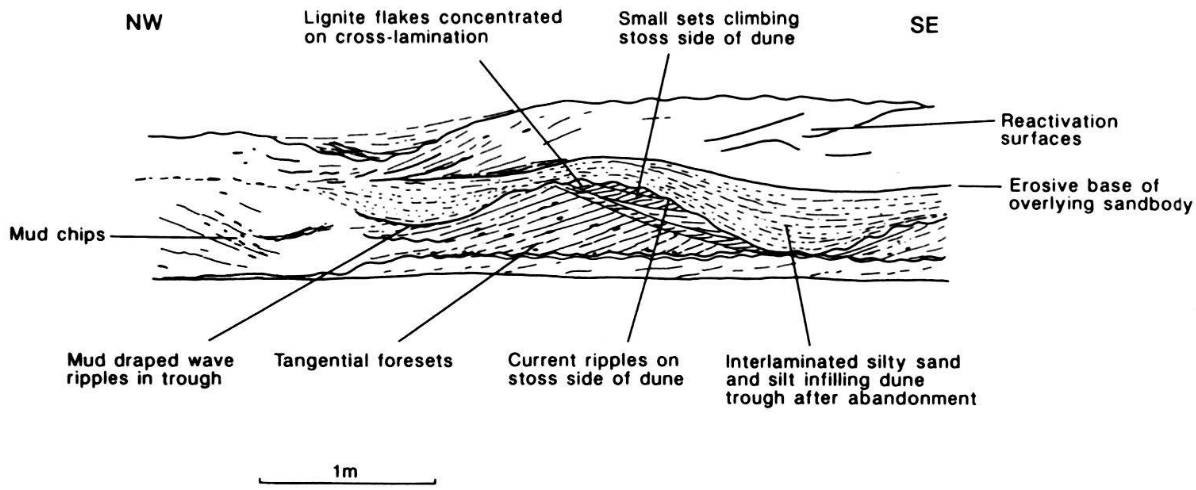


Fig. 22. Detail of trough cross-stratified sand facies in the Grésy lithosome, showing the preserved dune profile. From the "Ponliniere" stream section at 865.5/076.8, 1 km N of the village of St. Pierre-d'Alvey. From field sketch with same vertical and horizontal scales.

ment transport rates were lower and flow vortices over bedform slipfaces were reduced. The scour-pit fills were preserved by burial resulting from subsequent bedform climb during springs, or by abandonment preserving dune form-sets. Wave ripple crestlines in this facies are orientated N – S (Fig. 21 B), suggesting that they have been refracted into rough parallelism with the regional trend of the eastern shoreline of the seaway. Palaeocurrents from trough cross-strata are widely dispersed and polymodal, with inferred migration directions to the E, S and SW.

3. **Low-Angle Planar Cross-Stratified Sandstones,** composed of clean, well-sorted very fine to fine grained (fine to medium in the Pont-de-Beauvoisin lithosome) sandstones in units from 1 m to 10 m thick infilling very broad, shallow depressions. Low-angle ($< 5°$) cross-strata locally show minor scouring and low-angle truncations. Stratification is also picked out by muddy partings or flasers and locally by wave ripple-marks. This facies commonly erosively overlies the large-scale cross-stratified sandstones and trough cross-stratified sandstones.
The low-angle parallel-lamination suggests deposition as upper stage plane beds. The fine-grained intercalations indicate periods of much lower energy. The occurrence above tidal sandwaves and dunes suggests deposition on emergent shoals or filled channels. The parallel-lamination may have been produced under the high energies of shoaling waves in very shallow water.

4. **Horizontally-Stratified Sandstones** (found only in the Grésy lithosome) in units between 1 m and 20 m thick with cm- to dm-scale laterally continuous beds composed of very fine sandstone. Layers of parallel horizontal lamination occasionally have wave-rippled tops. Lamination surfaces have parting lineations and

Fig. 23. Eight detailed logs through the interbedded sands and silts of the Grésy lithosome in the le Sierre valley, about 3.5 km NE of Grésy-sur-Aix on the D911 road (882.8/088.5). Sedimentary cycles are mostly thinning-up and fining-up. A distinct erosive surface is shown in the dashed line. The detail shows thin cycles of rippled sands passing up into pinstripe lamination.


laminated silt
linsen
flaser
pinstripe lamination
wave ripple
E
occasional cross-lamination visible
planar erosion surface
small wave ripple forming in the lee of previous wave ripple
10cm
LOG 1
LOG 2
LOG 3
LOG 4
LOG 5
LOG 6
LOG 7
LOG 8
Height in m
0
0.2
0.4
0.6
0.8
1.0
1.2
1.4
1.6
1.8
2.0
2.2
2.4
2.6
2.8
5m
orientation of symmetrical ripple-mark crestline
orientation and sense of asymmetry of crestline

corrasion ripples (microripples; Allen 1984). Undulatory lamination locally fills small erosive scours.
The facies was deposited primarily as upper stage plane beds. Periodic exposure is indicated by the corrasion ripples formed by wind scouring of the wet very fine sand surface (Allen 1984; Fedo & Cooper 1990). Periods of cessation of upper flow regime flow are indicated by the wave rippled top surfaces of beds. An origin in intertidal sandflats or their seaward-facing edges is plausible. The N–S orientation of wave ripple crestlines (Fig. 21 B) suggests that waves had refracted in shallow water on the west-facing coast.

5. **Massive Sandstones** (found only in the Grésy lithosome) in units 1 m to 30 m thick composed of bioturbated very fine sandstone with a mottled appearance caused by dispersed flakes of lignite (Fig. 19a). Slow accumulation may have favoured the complete reworking by organisms of these sandstones.

4.5.2 Heterolithic and Fine-Grained Facies

Three heterolithic and fine-grained facies have been recognized:

1. **Interbedded Sandstones and Siltstones** in units 2 m–40 m thick of great lateral continuity (many tens of m) (Fig. 23). The top surfaces of siltstone laminations are commonly ornamented with sinuous feeding trails (eg. *Didymaulichnus*; Hakes 1976, and the biloped *Palaeobullia*; Knox & Miller 1985). The facies is made up of cycles up to 0.5 m in thickness: (i) Asymmetrical fining-up/thinning-up packages (Fig. 23) with erosive bases over other FU–TU cycles. (ii) Symmetrical cycles which thicken and coarsen up, then thin and fine up (only found in the Grésy lithosome) with non-erosional boundaries.
The cyclical deposition of sandstone-siltstone couplets indicates deposition as bundles under tides varying in strength through the neap-spring cycle. The asymmetrical cycles preserve only the spring to neap part of the cycle, whereas the symmetrical cycles preserve the entire lunar cycle. The neap-spring cyclicity is clearly picked out in bed thickness plots (Fig. 24a). Tessier and Gigot (1989) and Tessier et al. (1989) also describe the near-perfect preservation of neap-spring cyclicity in vertically accreted bundles in heterolithic deposits. Sediment transport during dominant tides was in the form of migration of current ripples, but with a wave component, particularly seen by symmetrical rippling at the end of the tidal flow. This facies passes laterally in some places into the toes of tidal sandwaves, suggesting that it is subtidal in origin. Wave ripple crestlines are orientated roughly E–W (Fig. 21 b), suggesting that the formative waves propagated longitudinally along the seaway, more or less unaffected by the regional trend of the coastline.

2. **Linsen and Pinstripe Facies** (in the Grésy lithosome) consisting of intercalated very fine sandstone and sandy siltstone. It is composed of rhythms of one or two sandstone or silty sandstone layers draped by one or two sandy siltstone layers (Fig. 25). The sandstones typically contain parallel laminations, climbing ripple cosets, and climbing wave ripples with trochoidal top profiles.
Both the thickness of the rhythms and their internal structural composition vary through cycles (Fig. 24b). In the thicker rhythms, the deposits of both the domi-

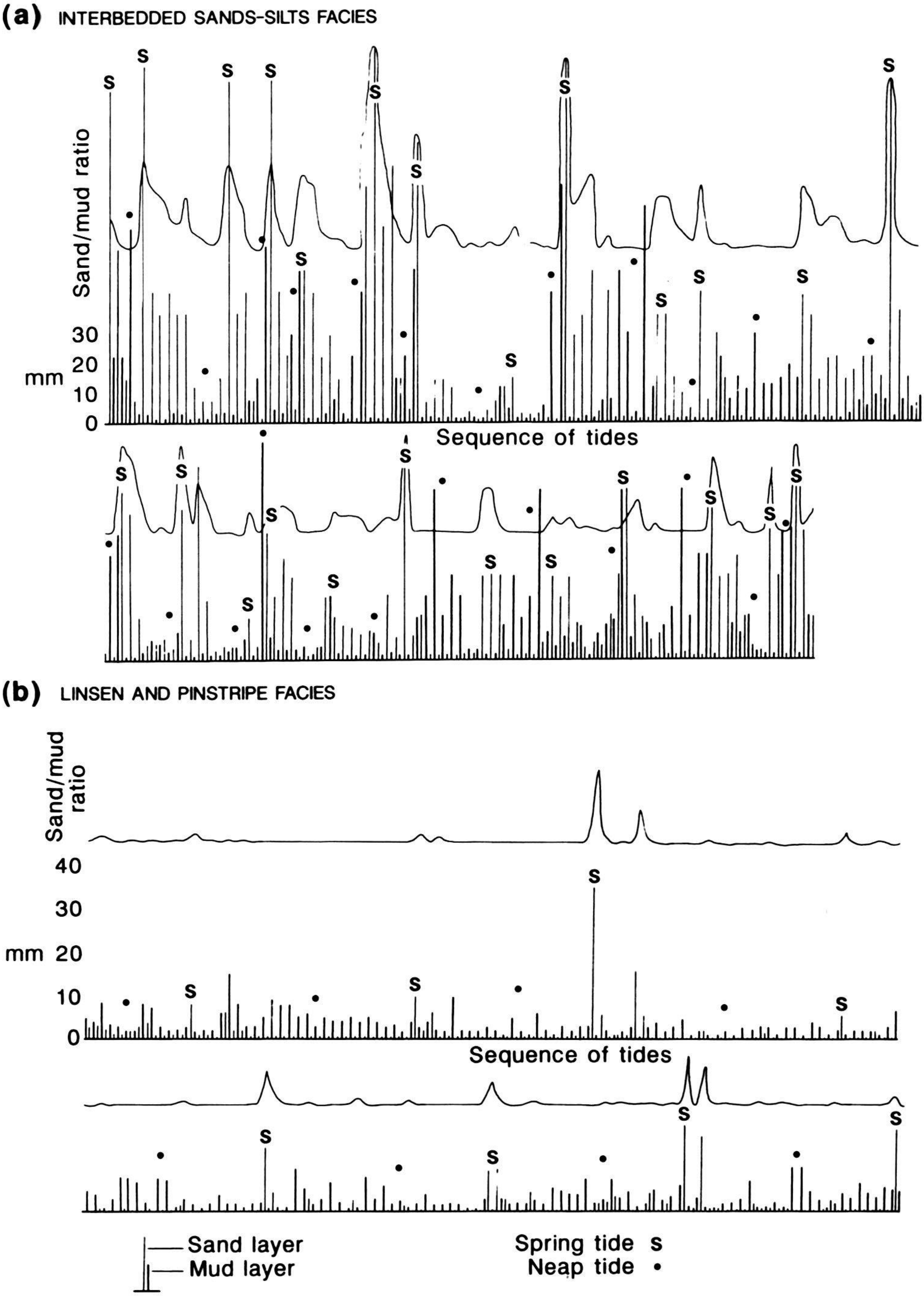


Fig. 24. Cyclicity in Grésy lithosome heterolithics. (a) Measured sequence of sand and silt couplets in the interbedded sand and silt facies, with corresponding sand/mud ratio. Measured in the Ponliniere stream section at 865.3/076.8, 1 km north of St. Pierre-d'Alvey. (b) Measured sequence of sand and sandy silt couplets from the linsen and pinstripe facies, with corresponding sand/mud ratio. From the Merdaret stream section at 866.1/078.9, 0.5 km NW of le Verdan.

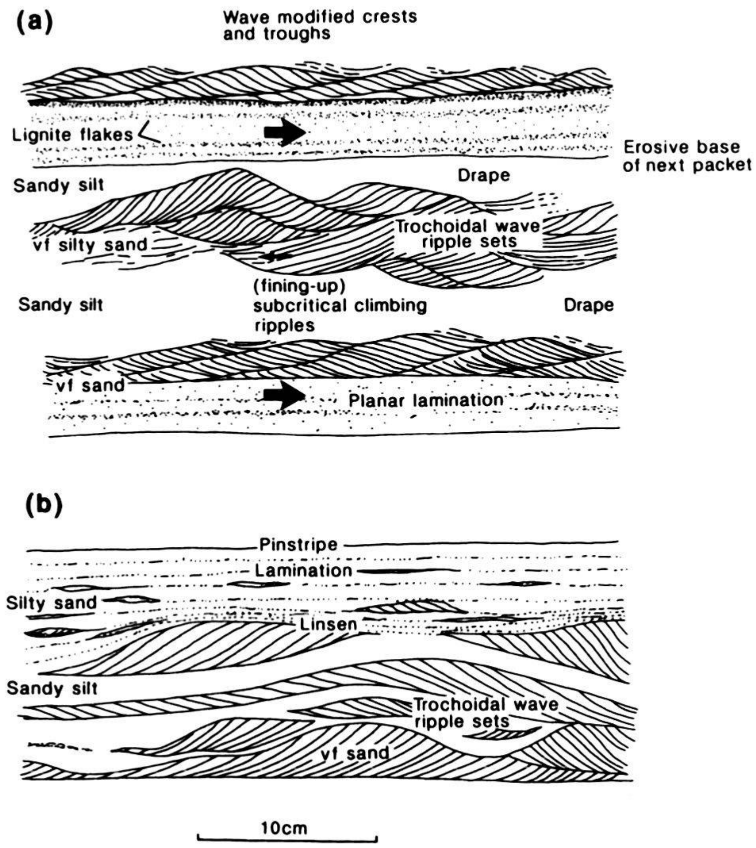


Fig. 25. Details of small-scale sedimentary structures in Grésy lithosome heterolithics. (a) A sedimentary packet from the linsen and pinstripe facies consisting of two sand/sandy silt couplets deposited during one ebb-flood cycle. The dominant tide couplet consists of planar horizontal lamination overlain by subcritically climbing current ripples and a sandy silt drape. The subordinate tide couplet consists of interwoven ripple cross-laminated sets modified by waves, followed by a sandy silt drape. From the Merdaret stream section at 866.1/078.9, 0.5 km NW of le Verdan. (b) Detail from the linsen and pinstripe facies showing swollen rippled lenses of sand and sandy silt drapes passing upwards into starved ripple lenses and finally pinstripes of very fine sand in a background of sandy silt. The upward fining and upward decrease in volume of sediment deposited may reflect hydrodynamic changes through a spring to neap cycle. From the Ponliniere stream section at 865.1/076.7, 1 km NW of St. Pierre-d'Alvey. From field sketches with same horizontal and vertical scales.

nant (plane beds and ripples) and subordinate tide (ripples) are preserved, together with the fine sediment drapes of slack water periods. In the thinner rhythms, subordinate current bedforms are absent and drapes are single. This suggests that the facies was deposited close to the base of the intertidal zone on fine grained rippled flats which accreted under a neap-spring control. Wave ripple crestlines trend NW–SE (Fig. 21 b), but are likely to reflect the local propagation direction of waves in a zone of complex coastal topography.

3. **Laminated Siltstones** in units between 1 m and 8 m in thickness, composed entirely of grey siltstone or clay, with a small proportion of very fine silty sandstone. Millimetre-scale laminae, lenses and starved ripples occur, and lamination surfaces are typically covered by sinuous grazing trails. A quiet water environment such as a muddy tidal flat or a protected lagoon is envisaged.

4.5.3 Other Facies

A number of miscellaneous, but important, facies also occur.

Channel-fills (Grésy lithosome), generally about 1 m thick, steeply incise into adjacent fine sandstones and siltstones, and are composed of either concordantly stratified heterolithics, or low-angle accretionary sandy wedges with mud pebbles. They are thought to have formed by the filling of intertidal creeks. One example of a thicker (9 m) channel-fill is filled by a chaotic mass of rounded mudstone and sandstone boulders, and clearly excavated to subtidal depths, before being filled by the products of bank collapse. The channel-fills are associated with ubiquitous soft-sediment deformation.

Pebbly and Shelly Lags (Grésy lithosome) unassociated with channels, contain extraformational debris (sandstones, bored limestones, chert) and reworked bioclastic material (oysters, fish teeth, wood fragments). It is likely that these units (0.2–0.3 m thick) have been winnowed *in-situ*. Together with the occurrence of deeper water facies above the lags, this suggests that they formed by transgressive shoreface retreat (ravinement of Stamp 1921; Bruun 1962; Nummedal & Swift 1987).

Conglomerates, such as the **Cognin Member**, which is only found on the western flank of the Chambéry Syncline, is a 24 m-thick unit of well rounded, well sorted (2–3 cm diameter), well stratified fine conglomerate composed of grey limestone, quartzite, veined limestone and red chert. Wave reworking of pebbles transported along the eastern shore of the seaway by longshore drift from a sediment source in the south (Voreppe area) seems the most likely explanation for the Cognin Member. The thicker **Chamoux Conglomerate** has not been investigated in detail. It contains crude, sub-horizontal stratification with normal and inverse grading, locally-developed imbrication and sandstone wedges infilling scoured hollows in the underlying conglomerates. Pebbles and cobbles are rounded to subrounded, and include a wide variety of rock types – dominantly bored limestone, but also radiolarian chert, quartzite, and fine grained igneous rocks, but high-grade metamorphics are rare. Deposition in a marine fan-delta complex, dumping pebbles and cobbles directly into the nearshore zone where they were they were bored by bivalves before burial, is possible. Imbricated pebbles indicate an east to west current direction.

4.5.4 Stratigraphical and Palaeogeographical Summary

The distribution of the various facies in the Grésy lithosome provides important information on stratigraphic trends. The intertidal facies predominate low in the section in the east, but occur higher in the section as one moves westward. Subtidal facies, on the other hand, are more abundant in the lower halves of all logged sections, and especially those in the west. This suggests that the coastline prograded westwards through time, although there are many high-frequency fluctuations in coastal position superimposed on this progressive trend. Vertically stacked groups of facies indicate gradual shallowing up, followed by rapid deepening, so that the Grésy can be thought of as a set of seaward stepping genetic sequences or cycles. The palaeogeography in Grésy times is therefore one of an extensive tidal flat system on the eastern side of the seaway, with a wave-worked front, subtidal to intertidal shoals and subtidal channels, as envisaged by Lamiraux (1977). The macro-tidal flats around the Wash, eastern England (Evans 1967), the meso-tidal estuaries of the east coast of USA (Greer 1975), the tidal flats and inlets of the inner part of the German Bay (Reineck 1972; Reineck & Singh 1980) and the meso-tidal subtidal and intertidal environments of Willapa Bay, Washington, USA (Clifton et al. 1989) may be appropriate modern analogues.

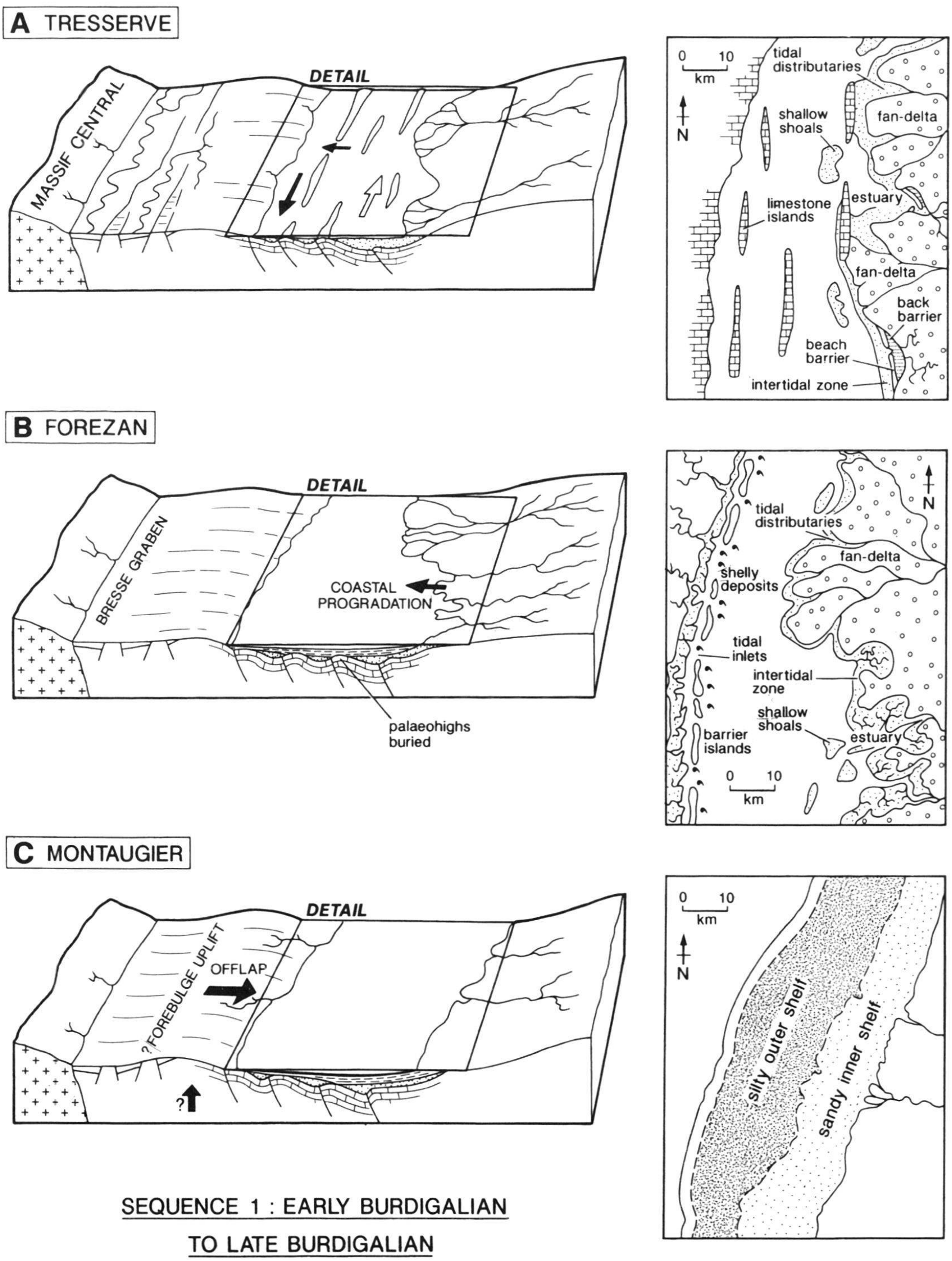
The great thickness (670 m in the Novalaise Syncline) of the Grésy lithosome (late Burdigalian – early Langhian) testifies to high sediment accumulation rates and therefore high delivery rates of sediment from the Alpine hinterland. Since the Grésy lithosome is found in the Rhône valley and Lyonnais region, the seaway must have expanded markedly at this time, reaching a minimum width of 80 km.

Eventually, in Pont-de-Beauvoisin times, the region was occupied by a coarse grained marine fan delta (Chamoux Conglomerate) which dominated the Chambéry Syncline area, and, to the west, high-energy tidal nearshore sands of the Pont-de-Beauvoisin lithosome. This palaeogeography is similar to that envisaged for the Alpine coast of the seaway elsewhere in the Molasse Basin in Switzerland during Burdigalian times (Hofmann 1960; Lemcke 1973; Matter et al. 1980; Homewood & Allen 1981; Allen et al. 1985; Homewood et al. 1986; Lejay 1991 unpubl.). During the Langhian-Serravallian the eastern shore prograded rapidly to the west, depositing the Chamoux Conglomerate. The coarse-grained fan-delta system continued to prograde during the Serravallian, with the deposition of the (?)lower Tortonian shallow marine "*Sables de Chimilin*" west of the study area in the northern Bas-Dauphiné (Lamiraux 1977; Nicolet 1979). This was followed by widespread regression and the deposition of the Upper Freshwater Molasse, occluding the marine seaway from the Alpine perimeter all the way from eastern Switzerland (Bürgisser 1980) to the Rhodano-provençal gulf of southern France (Valensole Conglomerate; Clauzon et al. 1987).

5. Implications for Basin Evolution

The Tertiary stratigraphy of the Rhône-Alp region can be divided into two main stratigraphic sequences *sensu lato* which have, at their base, major surfaces of marine onlap. This is not to say that a further subdivision into chronostratigraphically-significant packages of genetically-related strata (ie. depositional sequences *sensu stricto*) is not possible. However, the outcrop conditions and poor biostratigraphical control do not justify this at present, particularly because we are unable to fully evaluate the local, regional or interregional importance of erosional surfaces between lithosomes. Furthermore, we do not wish to imply any false precision in the numerical ages of the stratigraphic sequences. Very few absolute age dates are available. Correlation with the high frequency excursions of the Cenozoic segment of the so-called "global" cycle chart of Haq et al. (1987; 1988) has therefore been avoided, despite the fact that Neogene stratotypes were apparently obtained primarily from western Europe.

Fig. 26. Partly speculative palaeogeographical evolution of the Rhône-Alp region through Sequence 1. Sequence 1 is marked by the initial flooding of the region, the development of a progradational tide-dominated coast in the east, onlap of the western basin limit, ending with a period of muddy shelf sedimentation, signifying a basin deepening and shut-down in sediment supply. This was accompanied (or immediately followed) by stratigraphic offlap at the distal, feather edge of the basin.


A TRESSERVE
MASSIF CENTRAL
DETAIL
0 10 km
N
tidal distributaries
shallow shoals
fan-delta
limestone islands
estuary
fan-delta
back barrier
beach barrier
intertidal zone
B FOREZAN
BRESSE GRABEN
DETAIL
COASTAL PROGRADATION
palaeohighs buried
N
tidal distributaries
fan-delta
shelly deposits
tidal inlets
intertidal zone
shallow shoals
estuary
barrier islands
0 10 km
C MONTAUGIER
DETAIL
?FOREBULGE UPLIFT
OFFLAP
?
0 10 km
N
silty outer shelf
sandy inner shelf
SEQUENCE 1 : EARLY BURDIGALIAN
TO LATE BURDIGALIAN

5.1 Sequence 1 – early to late Burdigalian (c. 21–23Ma to 16.5–17.5Ma)

The unconformity at the base of the Burdigalian represents a major transgressive surface. The surface was initially covered with large-scale tidal sandwaves, before a sandy wedge (Tresserve lithosome) prograded from the Alpine margin (Fig. 26A). The Tresserve lithosome baselaps onto Aquitanian sediments and Mesozoic basement and pinches out at a position close to Mont Tournier, thus defining the western margin of the basin at this time. The eastern margin of the basin has since been obscured by overthrusting in the Subalpine chain. The horizontal distance between Mont Tournier and the Subalpine thrust front east of Aix-les-Bains is 18 km (Fig. 5). Allowing for shortening in the basin, the minimum E–W width of the Tresserve is therefore about 20 km. There is thought to have been about 15 km displacement on the frontal thrust of the Subalpine chains in the Chartreuse section (Mugnier et al. 1990), and the Chartreuse and Bauges Massifs, which contain remnants of lower Miocene sediments, have themselves been internally shortened (Mugnier et al. 1990; Guellec et al. 1989, 1990). It is certain, therefore, that the seaway was considerably wider in the early Miocene (in excess of 35–40 km) than the distribution of present-day outcrops initially suggest.

Condensed shelly limestones of the Forezan lithosome overlie basement on the eastern flank of the Bas-Dauphiné at the stratigraphic pinch-out (Fig. 26B), representing a further 5 km of marine onlap. Since the deformation propagated into the Jura province through the Miocene (Mugnier & Ménard 1986), the westward advance of the Alpine limit of the basin may have kept pace with or even outpaced the westward onlap of the feather edge, so it is extremely difficult to estimate changes in the width of the seaway during this time interval.

The Montaugier lithosome marks a period of stratigraphic offlap on the western side of the basin, and deepening of water depths/starvation of the basin (Fig. 26C). The amount of offlap is 7 km at the western margin. At the very least, therefore, the basin narrowed, unless the Alpine front was profoundly flooded at the same time.

5.2 Sequence 2 – latest Burdigalian to Serravallian (16.5–17.5 Ma to 12–15 Ma)

The base of the Grésy lithosome marks a period of clastic influx into the seaway and renewed marine onlap on the western side of the basin (Fig. 27D). The western margin lay in the Bas-Dauphiné, and the depocentre migrated in concert into the Novalaise Syncline. The flooding extended beyond the boreholes at La Tour-du-Pin, where the Grésy lithosome directly overlies continental Oligocene sediments, to the stable margin of the Massif Central in the Rhône valley (Demarcq 1984). This created a considerably expanded seaway. The Pont-de-Beauvoisin lithosome records the overall coarsening of the succession and the progradation of coarse grained fan-deltas derived from the Alps in the form of the Chamoux Conglomerate (Fig. 27E). Whereas the Alpine margin presumably continued to encroach into the peripheral foreland basin, the outboard margin was pinned against the stable Massif Central at this time. This would have narrowed the seaway during the Serravallian.

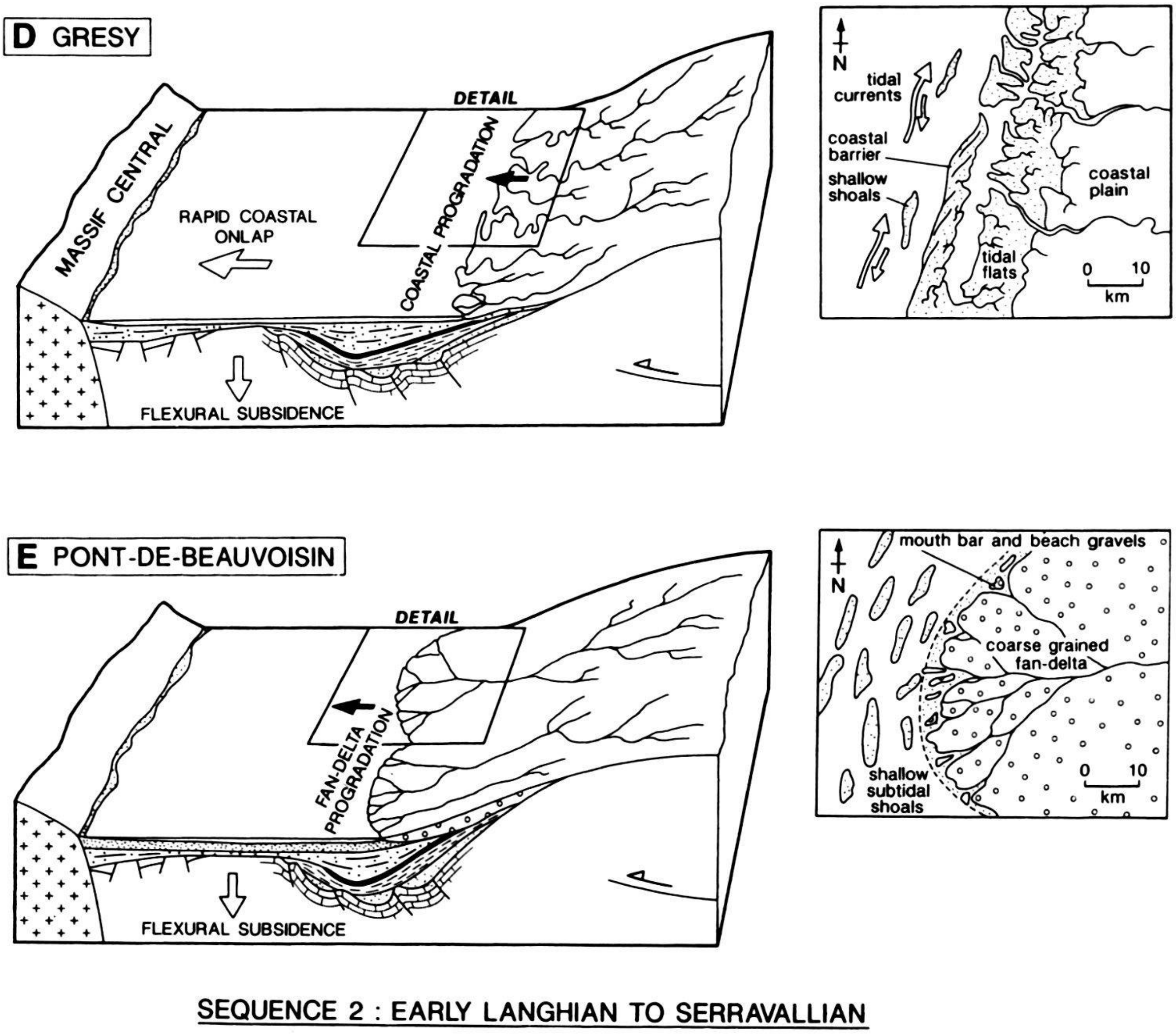


Fig. 27. Partly speculative palaeogeographical evolution of the Rhône-Alp region through Sequence 2. The Grésy lithosome strongly onlaps or downlaps the distal basin margin, spreading tidal deposits as far as the Massif Central at the western edge of the Valence basin. The fine grained tide-dominated coastline on the Alpine margin was later replaced by coarse grained fan-deltas feeding sediment into the tide-swept shallow sandy sea of the Pont-de-Beauvoisin lithosome.

5.3 Discussion

Four marine depositional sequences *s.s.* have been recognized in the Miocene of the Rhône valley (Gourinard et al. 1985; Anglada et al. 1988; Lesueur et al. 1989), and bounding unconformities have been correlated with the Haq et al. (1987) curve (TB2.1 to TB2.4 cycles). These depositional sequences have also been identified in the Miocene of the foreland basin of the Digne area (Crumeyrolle et al. 1991). The Upper Marine Molasse has long been known to consist of two lithostratigraphic divisions in the Swiss Molasse basin and in western Austria (the "Burdigalien" and "Helvetien" of Heim et al. (1928); the Luzern Formation and the St. Gallen Formation of Keller (1989) and Schaad et al. (1992)). These two divisions are separated by a basin-wide transgressive surface at c. 19 Ma (Keller 1990). The two divisions of the molasse of eastern Switzerland and western Austria and the two "sequences" of the Rhône-Alp may therefore be equivalents, though there are considerable difficulties in extrapolating from region to region where lithostratigraphical boundaries may be highly diachronous.

The mechanisms driving the basin evolution during the Miocene in the Rhône-Alp region are not fully constrained. Rather than broadly speculate, we discuss here a small number of topics which serve as pointers to future studies.

5.3.1 Significance of the Basal Unconformity

From a broad region around the Alpine perimeter there is evidence of marine transgression at the onset of the Burdigalian stage of the early Miocene. Indeed, the early Burdigalian is shown as a period of rapidly rising sea level on the "global" cycle chart of Haq et al. (1987) (TB2.1 stage). However, it has recently been shown that the Burdigalian transgression in the Molasse basin of central Switzerland may be due, at least in part, to tectonic events in the mountain belt (Sinclair et al. 1991). These authors associated uplift on the basinward-facing flank of a flexural forebulge, relative deepening in the basin, and a slowing of the thrust front advance rate to a rejuvenation of the supracrustal load at the rear of the orogenic wedge. Modelling suggests that this has the effect of "dragging" the forebulge towards the mountain belt and deepening the basin close to the thrust front. Subsequent erosion combined with renewed tectonic advance could explain the stratigraphic onlap onto the forebulge during the Burdigalian without recourse to eustatic change. In the western Alps, Guellec et al. (1990, p 169) also recognize the start of the Burdigalian as a time between a late Oligocene-Aquitanian phase of thrusting, forming the internal parts of the Subalpine chains, and a subsequent Miocene-Pliocene phase which detached the basin. This also suggests that the basal Burdigalian unconformity may have a tectonic origin, at least in part.

5.3.2 Significance of late Burdigalian Offlap

The association of offlap at the distal feather edge of the basin and sediment starvation within the basin has been used as evidence characteristic of viscoelastic relaxation of the lithosphere during times of tectonic quiescence (Beaumont 1981; Quinlan & Beaumont 1984; Tankard 1986; Beaumont et al. 1988). This is a possibility for the "Montaugier" event in the Rhône-Alp region. Another possibility is of internal rejuvenation of the orogenic wedge on a weak but elastic plate, as envisaged in the central Alps (Sinclair et al. 1991). A third possibility is of eustatic sea level change. Since the available deformation history (eg. Mugnier & Ménard 1986; Guellec et al. 1990) does not have the necessary chronological refinement, it is difficult to test the first two possibilities. Problems in biostratigraphic correlation also make the third possibility difficult to test at present.

5.3.3 Role of Early Rifting on Foreland Basin Evolution

The North Alpine Foreland Basin and its French segment was superimposed on a lithosphere that had already been stretched during the Permo-Carboniferous

(Ménard & Molnar 1988), then thinned again during the Mesozoic development of Tethys and its passive margin (Trümpy 1980).

The western European rift system initiated on these ancestral, mostly NNE–SSW orientated faults in the late Eocene (Rigassi 1977b; Bergerat 1987; Bergerat et al. 1990). These opened up under E–W extension during the Oligocene, producing a series of salt basins from the southern Rhine graben to the southern Rhône valley (Rat 1978; Debrand-Passard & Courbouleix 1984; Ziegler 1988). In particular, the Bresse graben structures extended southwards into the Rhône-Alp region (Bergerat 1987). The normal faults cutting basement and Palaeogene cover in the Rhône-Alp region have also been explained as due to outer-arc extension associated with flexure of the European plate (Mugnier & Ménard 1986) in much the same manner as recently suggested by Bradley & Kidd (1991). From the late Oligocene, WNW-directed convergence (Dewey et al. 1973; Ricou & Siddans 1986; Gillcrist et al. 1987; Vialon et al. 1989) incorporated Helvetic flysch and Lower Freshwater Molasse (Chattian-Aquitanian) into the Subalpine thrust sheets (Masson et al. 1980; Doudoux et al. 1982; Tardy & Doudoux 1984). In the Subalpine massifs, the Burdigalian OMM is found locally unconformably overlying deformed Aquitanian molasse, suggesting that the latter was deposited in piggy-back basins during active shortening (Mugnier & Ménard 1986). In a more external position in the Rhône-Alp region, the NNE–SSW trending (?normal) faults affect Oligocene to Aquitanian deposition, and also affect marine facies and sediment dispersal in the Burdigalian. Whether by this stage the faults had already undergone inversion during Alpine compression is not known. By the late Burdigalian the fault-related palaeohighs had been buried by the marine sediments of the peri-Alpine seaway. This is clearly a very different scenario to that envisaged for the North Alpine Foreland Basin in Switzerland (Pfiffner 1986; Homewood et al. 1986; Sinclair et al. 1991; Allen et al. 1991), where flexural subsidence dominated other effects from the late Eocene onwards. The Rhône-Alp region therefore occupies a pivotal position between the Rhine-Bresse-Rhône rift system and the peripheral foredeep of the Alps.

6. Conclusions

1. The Upper Marine Molasse (OMM) in the Rhône-Alp region spans the time interval from early Burdigalian to Serravallian, a period of about 10 My entirely within the Miocene. The maximum thickness preserved is 1150 m, representing a sediment accumulation rate of just over 0.1 mm y^{-1} ignoring post-Miocene compaction.

2. The early Burdigalian seaway was established by the flooding from the south of a number of narrow N–S or NNE–SSW orientated continental basins related to late Eocene-Oligocene extension in the Rhine-Bresse-Rhône system. The oldest deposits are large scale subtidal sandwaves which migrated towards the south and SW, that is, towards the entrance of the tidal strait. Subsequently, a tide-dominated coastal tract prograded westwards into the basin from the Alpine flank, while condensed shelly limestones accumulated against a quiet rocky shore in the west. By the mid-Burdigalian the N–S orientated palaeohighs had been buried by marine deposits of the peri-Alpine seaway. The mid-late Burdigalian was a period of low-energy conditions associated with a muddy shelf that became sandier towards the eastern, Alpine coast. The remainder of

the Miocene into the Serravallian was characterized by the progradation of thick sandy wedges from the Alpine mountain belt into the seaway. An initially fine-grained tidal coastline was replaced by a conglomeratic marine fan-delta as the entire basin was translated westwards to the edge of the Massif Central.

3. The Miocene stratigraphy can be divided into two main stratigraphic sequences (*s. l.*), each with a major surface of marine onlap at its base. The first sequence, from early to late Burdigalian (21–23 Ma to 16.5–17.5 Ma), is found in the east of the study area, principally in the Chambéry, Rumilly and Novalaise Synclines. The top of the sequence (late Burdigalian) is marked by muddy shelf conditions and 7 km of offlap at the western margin of the basin. This is thought to have been associated with basin narrowing. The second sequence, from latest Burdigalian/early Langhian to Serravallian (16.5–17.5 Ma to 12–15 Ma) is recognized by rapid marine onlap towards the west of tidal deposits over Mesozoic basement, the western edge of the basin reaching as far west as the stable margin of the Massif Central. Encroachment of the sediment-nourished Alpine margin caused a major coarsening-up trend to develop. The precise linkage of stratigraphic history to events in the orogenic wedge has not yet been developed.

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