Ghent, 10 – 13th September 2001

International Conference

ORIGIN, PROCESS, AND EFFECTS OF

SUBSURFACE SEDIMENT MOBILISATION

ON RESERVOIR TO REGIONAL SCALE

POST-MEETING FIELD TRIP IN THE VOCONTIAN BASIN (SE-FRANCE)

(14-15th September, 2001)

organized by
Olivier PARIZE

Field trip guide book

DES ASSOCIATION SEDIMENTOLOGISTES FRANÇAIS

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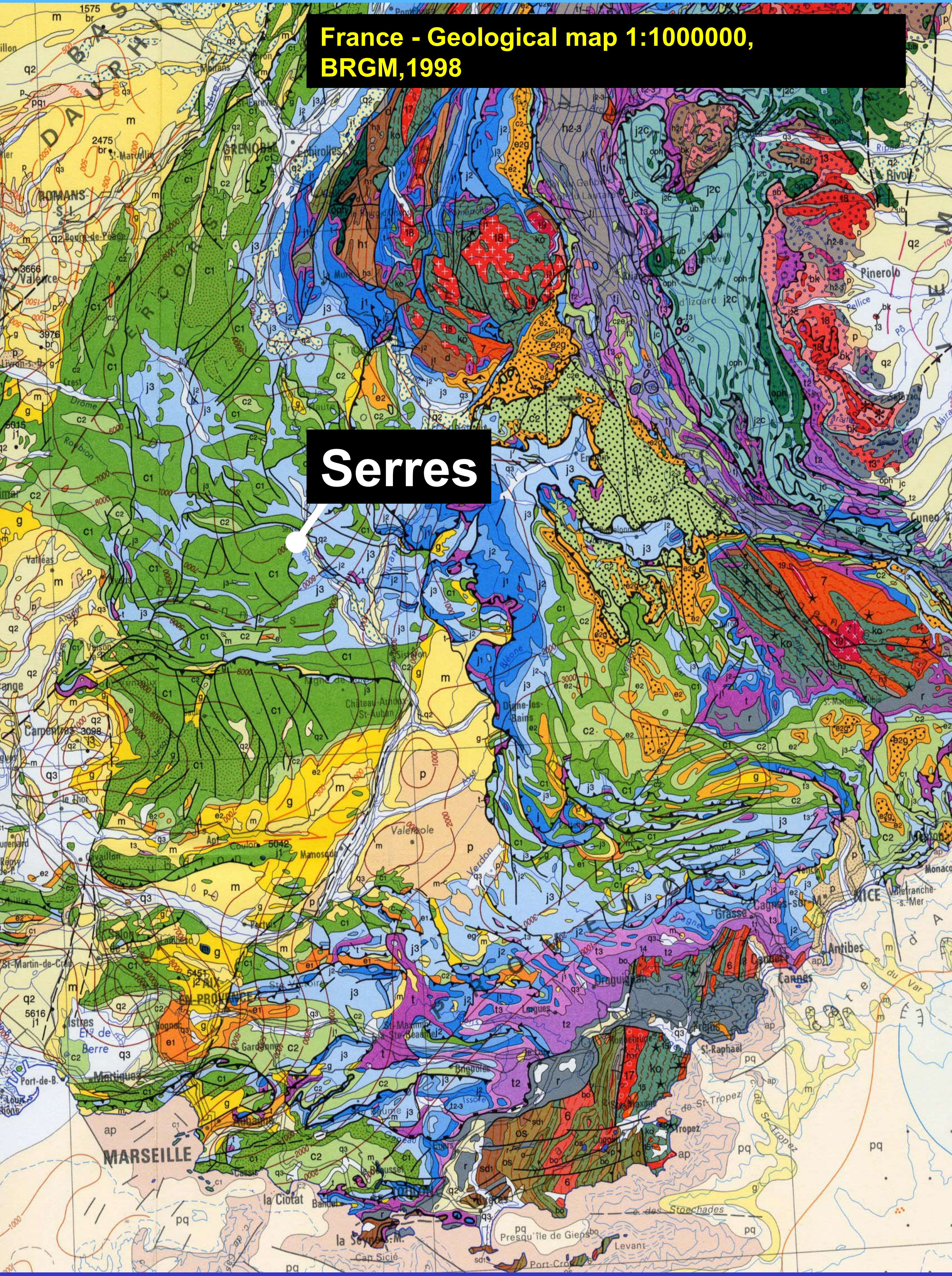
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Guide Book Field Trip #2,
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ECOLE DES MINES
DE PARIS

14-15th September, 2001

SAND INJECTIONS IN DEEP MARINE SHALE FORMATIONS IN SE-FRANCE

Organized by Olivier Parize

with collaboration of Gérard Friès
And Jean-Loup Rubino

TOTALFINAELF

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MINES

GÉOLOGIE SÉDIMENTAIRE

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35, rue Saint-Honoré 77305 FONTAINEBLEAU

2001 Ghent International Conference

SUBSURFACE SEDIMENT CONFERENCE

POST-MEETING FIELD TRIP

IN THE VOCONTIAN BASIN (SE FRANCE)

14th to 15th of September 2001

SAND INJECTIONS

IN DEEP-MARINE SHALE FORMATIONS

Examples from Vocontian outcrops

(SE BASIN, FRANCE)

Key words:

Dyke, dike, sill, clastic dyke, sandstone dyke, sandy injection, turbidite, compaction, outcrop, Alps, France, Cretaceous, passive margin, slump, debris flow, massive sand,

Olivier Parize — **Ecole des Mines de Paris**

With Gérard Friès — **and** — **Jean-Loup Rubino**

TFE GRC London — **TFE, CST JF Pau**

Special thanks to Bernard Beaudoin, Sylvain Eckert, Michel Tijani (Ecole des Mines de Paris, Patrice Imbert (TFE) and Frédéric Schneider (Institut Français du Pétrole)

TECHNICAL PROGRAMME

Leaders :

O. PARIZE, Ecole des Mines, Fontainebleau

And

G. FRIES & JL RUBINO TOTALFINAELF

Day 0, Thursday 13th September:

Transfert from Ghent to Serres

Meet at train station Ghent St Pierre at 17:00

Transfert from Ghent to Brussel airport.

Travel from Brussel to Marseille-Provence airport by plane:

Flight SN3603 Brussel – Marseille at 19:30

Arr in Marseille-Provence airport at 21:20

Drive from Marignane (Marseilles-Provence Airport) to Serres

Diner and night in Fifi Moulin Hotel, Serres

The Vocontian domain: general presentation (talk)

Some keys to better visit the Vocontian domain (main stratigraphical markers and main morphological features). The Vocontian domain provides a good example of confined setting located on the upper slope of the northern margin of the Lower Cretaceous Tethys-Valaisan ocean.

Day 1, Friday 14th September:

Rosans Upper Aptian sandstones

Drive from Serres to Rocher de Beaumont (from 700 to 1500 m high, 15 kilometres).

Firstly, introduction to the Vocontian basin, structure and stratigraphy.

Deep-water massive sands, turbidites, debrites and slumps; hemipelagic and pelagic "background" sedimentation and key beds.

Whole day out in the field near Saint-André-de-Rosans to observe a sill-dominated clastic injectite network: most of the day will be spent driving and walking on and around one massive sandbody and associated injectites: the Upper Aptian Rosans turbidite complex.

Injectites and injectite network geometries, facies, relations with the feeder, geometry of the *in situ* and injected complex, lateral continuity and reservoir geometry.

“Casse croûte provencal” on the outcrop

Diner and night in Fifi Moulin Hotel, Serres

Day 2, Saturday 15th September:

Bevons Upper Albian sandstones

Drive from Serres to Sisteron then Bevons (30 kilometres).

Whole day out in Bevons area near Sisteron to observe a clastic injection network and well exposed sills and dykes which can be up to 250m high. Their feeder is an Uppermost Albian massive channelised turbidite and located at the top of the hill. Detailed observations will be made of the patterns of sills, dykes and the source deposit in a dyke-dominated network. Injection facies, bed continuity, reservoir geometry, facies models.

“Casse croûte provencal” on the outcrop

Drive from Bevons to Marignane (Marseilles – Provence Airport at 18:00) via Aix en Provence TGV Station

Accommodation:

Hotel Fifi moulin
05700 SERRES – France
Tel. 33 (0)4 92 67 00 01
Fax 33 (0)4 92 67 07 56

You should remember that

your safety

is

your responsability.

If you feel uncomfortable with any of the stops,

then skip them.

SAFETY

On the two days, we'll climb up by about 200m. Nothing to worry about, but good hiking shoes are again recommended (plus plasters for those who take brand new shoes). Ties and formal dress optional.

Some localities visited during this field trip are roadside localities. Here the greatest danger is traffic. Stay alert, and be prepared to warn other members of the group of approaching vehicles.

The other principal danger relates to slopes. Some of the participants previous years found the trip a bit strenuous and thought some of the slopes were too steep: we can help you but it is essential you do only as much as you can and choose your own way up the slopes.

There are five localities where extra care needs to be taken:

Day 1: **Toulaye cliff, near Saint-André-de-Rosans,**
Serre d'Autruy, near Saint-André-de-Rosans,

Day 2: **Le Puy hill, near Bevons,**
Vieux Bevons hill, near Bevons,
Pierre-Avon, near Sisteron.

Care should be taken not to approach the lip of the cliffs and with the ascent and descent of the slopes. The group leaders will point out these hazards at the appropriate time in the field.

The group leaders will bring a basic first aid kit which will be carried with us.

FOREWORD

The topic of this field-trip is the clastic sills and dykes of the Vocontian domain of the SE Basin, France. The two spectacular outcrops, Bevons and Rosans, which will be visited during this field-trip, have been analysed firstly by B. Beaudoin and G. Friès at the beginning of 1980's.

One part of presented results has been obtained since 1983 in the frame of Ecole des Mines de Paris research programs: the "GeneBass" project with the support of CNRS, ELF and TOTAL (1983 – 1986) and specific sponsoring of ANDRA (1983 – 1986) carried out under the direction of B. Beaudoin, scientific and training collaborations with TOTAL, ELF then TotalFinaElf (since 1993) and scientific joint-venture "MinAndra" between ANDRA and Ecole des Mines de Paris (since 1999) carried out under the direction of B. Beaudoin and A. Trouiller.

R. Bouchet (1985), S. Eckert (2000), Ch. Cabrol (1985), Y. Grenetier (1984), S. Lalande (2000), B. Paternoster (1983), students at the Ecole des Mines de Paris, Y. Callec (1999), L. Dauphin (2002), C. El Maherssi (1992), J-N. Ferry (in progress), J. Maillart (1991), S. Mostefaï (1998), B. Pinoteau (1986), C. Rigollet (2001), F. Schneider (1988), V. Truyol (1991), PhD students at the Ecole des Mines de Paris, T. Bouchery (1998) and E. Salinas (1995) students at Lille university, have collaborated in the fieldwork as part of their respective degree courses. La Baume 3D mapping was conducted with F. Schneider during summer 1986. J-Y. Clément and M. Thiry have supervised the petrographic analysis of numerous X-Ray diagrams and thin-sections; B. Charlet, F. Hadj-Hassen, G. Vouille and M. Tijani, the mechanic modelling. We are extremely grateful to them for their friendly availability and their scientific support.

The detailed analysis of aptian-albian turbiditic systems of the Vocontian domain results from a research project (1993 – 1996) involving TOTALFINAELF (P. Imbert and J-L. Rubino) and ARMINES – Ecole des Mines de Paris (O. Parize). One of the results of this project is the "Vocontian model" to explain the location and the organization of deep-water massive sands. Clastic dykes and sills represent certainly the most particular facies of these channelled massive sands.

Some of these results have been previously published:

Sills et dykes gréseux sédimentaires :
Paléomorphologie, fracturation précoce, injection et compaction,
Thèse Géol., ENS mines de Paris, 1988
Mém. Sc. Terre, E. N. S. Mines de Paris, n° 7, 1988, 333 p.
Olivier Parize.

Sand injections in deep-marine shale formation in SE-France.
"Origin, process and effects of subsurface sediment mobilisation on reservoir to regional scale", post-meeting field-trip in the Vocontian basin (SE France),
14^{th} - 15^{th} September 2001
Mém. Assoc. Fr. Sédim., n. s. , n° 35, 124 p.
Olivier Parize.

Different works have been previously published about the stratigraphic, tectonic and sedimentologic settings of the Vocontian domain:

L'éventail sous-marin de Céüse (SE de la France) à l'Apto-Albien,
Guide book, field-trip 5^{th} and 6^{th} Oct. 1985,
Mém. Assoc. Fr. Sédim., n. s. , n° 2, 124 p.
Gérard Friès and Bernard Beaudoin.

Mesozoic eustacy record on western tethyan margins.
Post-meeting field trip in the vocontian trough,
Guide-book (25th - 28th november 1989),
Mém. Assoc. Fr. Sédim., n. s. , n° 12, 141 p.
Serge Ferry and Jean-Loup Rubino,

Dynamique du bassin subalpin méridional de l'Aptien au Cénomanien,
Thèse Doct. Sci., Paris, 1986
Mém. Sc. Terre, E. N. S. Mines de Paris, n° 4, 1987, 370 p.
Gérard Friès.

Phénomènes de resédimentation : Crétacé inférieur subalpin,
Guide book, field-trip n° 6, 51 p.,
5th Europ. Sedim. Meet., *Assoc. Fr. Sédim. and Int. Assoc.Sedim. Org.,*
Gérard Friès and Bernard Beaudoin.

L'Aptien et l'Albien de la Fosse vocontienne (des bordures au bassin). Evolution de la sédimentation et enseignements sur les événements anoxiques,
Thèse Doct. Sci., Tours, 1995,
Soc. Géol. Nord, publ. n° 25, 1997, 614p.
Jean Gabriel Bréhéret.

Notes

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INTRODUCTION

Vocontian Clastic Injections - September 2001

AIM OF THE FIELD TRIP

Sand injections are more and more described in deep-water sedimentary environment and 3D seismic analysis demonstrates the importance of their dimensions and the corresponding volume of sand involved in those phenomenons. They modify the geometry of deep-water reservoirs to a great extent (*e.g*. Jenssen *et al*., 1993; Newton & Flanagan, 1993; Shanmugam *et al*., 1994; Dixon *et al*., 1995). From subsurface data analysis, the injectite network appears to be located above the feeder; and the genetic model proposed by different authors therefore corresponds to a post-depositional injection implying the liquefaction of the sandy formation. This mechanism is supposed to occur under an important burial and should have been triggered by major earthquakes (Waterson, 1950; Marschalko, 1965) as proposed one century ago (Diller, 1889; Newsom, 1903) or by sudden overloading related to slump or debris flow deposit (Truswell, 1972; Hiscott, 1979; Shanmugam *et al*., 1994).
Several outcrops have been described as good analogues for sand injections: the Carboniferous of South Africa (Truswell, 1972) or Québec (Hiscott, 1979), the Jurassic and the Cretaceous of California (Diller, 1889; Newsom, 1903; Paterson, 1966; Smyers & Peterson, 1973) or Greenland (Surlyk, 1987, Leeds conference 1998; Gjelberg *et al*., London conference 2001), the Tertiary formation from the Californian coast (Newsom, 1903), the Carpathian mountains (Dzulinski & Radomski, 1953; Marchalko, 1965) or the perimediterranean Numidian flysch (Gottis, 1953; Colaccichi, 1959; Parize, 1988; Parize *et al*., 1999). However, the existing descriptions do not permit a fully 3D approach of the geometry of the injectite network, including the relations between the clastic sills and dykes and, firstly the feeder, secondly the timing of the injection and thirdly the sedimentary and structural context. Finally the current *per ascensum* and post-depositional model is not consistently supported by field observations (*e.g*. discussions *in* Waterson, 1950 or Truswell, 1972).
An alternative model, suggesting a syn-depositional injection during deposition of the channel feeder is proposed, based on the extensive studies of Bevons and Rosans outcrops in the Vocontian domain of S. E. Basin of France (Beaudoin & Friès, 1982; Beaudoin *et al*., 1983; Parize, 1988; Parize *et al*., 1999). The Bevons and Rosans areas are probably two of the most spectacular and well-exposed outcrops showing complex networks of clastic sills and dykes injected in a marly-limy host formation. At Le Puy hill near Bevons, where the vertical

and oblique injectite network is well developed, the dykes are mainly connected with the sills. Some minor dykes occur between the palaeo-sea floor and the main upper sill; however, the main vertical network is located stratigraphically under the sill network, located from 30 metres below the channel feeder and it penetrates the host formation down to 300 metres; at the Toulaye cliff near Rosans, the complete sills and dykes network and the channel feeder system are exposed along a 5 kilometres long continuous outcrop. The quality of these outcrops allows an accurate and detailed inventory of sills and dykes, so as a detailed study of these objects, and of their relations with the feeder body (Parize, 1988).

The purpose of this field trip is to present an original model, which includes the regional sedimentary and structural setting. We will present present in the next 4 days some geological facts (sedimentological and palaeomorphological setting, geometrical characterization of injectite network, injectite shape, relations with the feeder, time of the sandy injection, sand movement during injection) in order to propose a model explaining clastic injectites related to the Vocontian area based on a truly 3D geometrical approach and detailed field observations. The results of numerical simulations of the mechanical behaviour of a marly formation during its mechanical compaction and/or the sudden channelized infilling are presented to propose physically coherent explanations to the field observations and topics for debate.

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CLASTIC INJECTIONS,

SEPTA, NEPTUNIAN DYKES

AND MELANGES

DIFFERENT TYPES OF SAND INJECTITES

All objects which are oblique to the bedding (sills and dykes) *ie* exceptions to the "Superposition principle of stratigraphy" can be often gathered under the general heading of "sand injections". They can often be related to features of slope instability or post-depositional deformation features *e.g.* mud or sand volcanoes, polygonal faults in shale formations, pockmarks and gas hydrates features:

"Mélanges" or mixtures between sand and shaly deposits

Such mixtures of various origin most often occur during the depositional process of the different gravity-driven bodies. Several examples include:

A1 – The flow/deposition of the turbiditic current triggered a collapse/sliding of the destabilised flanks of the channel in two different ways: (i) erosion of the flanks and/or infill of asymmetric meandering channels; (ii) destabilisation and collapse of the flanks related to a wave effect of the flow (e.g. Stanley, 1975). This type of mixture is associated with specific channel infill.

A2 – The erosion at the base of high-density current triggered a lot of mud clasts ("slurry beds", e.g. Mutti & Normark, 1987, 1991).

A3 – Destabilisation and reworking of sandy and muddy deposits led to slumping and fluxoturbidites (e.g. Slaczka & Thompson, 1981).

These mixtures are also associated with sediment liquefaction and reflect soft-deformation prior to lithification (ie. Kneller and Mc Caffrey, 1999). They would be post-depositional features. In this case they look like pseudo-conglomerates with sandy elements floating in a marly matrix.

These injectite-form features are strictly restricted to a well-defined sedimentary body; they are part of it, like any other sedimentary structure. The corresponding sedimentary body might be described as a channel or channel/lobe deposit.

Sand injections in shale-dominated formations

These objects must have two (almost) parallel sides; their thickness is very small compared to the other dimensions. This definition is completed with the idea of superposition or intrusion: the veins do not belong to the host rock; they are secondary or epigenetic features. When analysing the injections, fracturation and injection have to be analysed separately; the sandy injection may occur in existing fractures or in fractures resulting from it. Finally the geometric definition does not necesseraly define the origin (syn- or post-depositional) of the vein.

Depending on its relation to the host formation, a vein which is oblique to the stratification, is called “dyke”, and one which is parallel, is called “sill”. Therefore, the clastic sills look like “beds”, which may be confused with real beds deposited onto the palaeo seafloor, like turbidites. However the sills differentiate from ordinary beds by a specific geometry in steps associated with the dykes.

The lithology of the fractured and injected host-formation will affect the general geometry of the injectites leading to less regular shapes than previously described:

B1 – In a “homogeneous” (isotropic) lithology like slumpings or debrites or superficial,wet sediment, fractures may present very complex and tortuous geometry.

B2 – In a well-stratified formation, the bedding anisotropy will cause a more regular fracturation pattern (before final compaction). The injectites are organised in regular sills and dykes.

The individual geometry of the sills and dykes is quite well-known. However, the sand body cannot completely be described without the global pattern of the injectites and their feeder reservoir. Several 2D geometric models have been proposed for the injectites (Dzulinski & Radomski, 1956; Truswell, 1972; Hiscott, 1979; Beaudoin *et al.*, 1983…), and for the relation between reservoir and sand injectites (Bielenstein & Charlesworth, 1965; Surlyk, 1987; Beaudoin *et al.*, 1985 …). 2D and 3D geometrical models linking reservoir and sand injectites

have been proposed at the Ecole des Mines de Paris (Beaudoin & Friès, 1982; Beaudoin *et al.,* 1983; Friès *et al*., 1984; Parize, 1988, 1998).
The individual and global geometry of the sand injectites appear to be constrained only by the mechanical behavior of the host-rock and not by the timing of the injection in the host (e.g. Surlyk, 1987, 1998 Leeds conference : two opposite models based on the same outcrops).

The injected sand is fed from a sand reservoir. Two main types of injections must be considered at that point: (i) syn-depositional and (ii) post depositional injections with respect to the period of deposition of the sand reservoir.

C1 – The small dykes in the vicinity of the channel related to shear during the infilling phase (eg. Fromaget key bed 1 at Toulaye – Day 3) and corresponding to neptunian dykes.

C2 –The post-depositional injection in the host-rock is the most frequently described mechanism in the literature (e.g. Maltman, 1994). The injection takes place under a thick overburden from under-compacted and over-pressured sandy lenses. Such a model was proposed very early by Diller (1889) and Newsom (1903); it implies non-sedimentary mechanisms (earthquakes) triggering sand liquefaction and fracturation of the host rock. This model explains correctly some “flower structures” described from subsurface data of the North Sea: Gryphon Field (Newman *et al*., 1993); Forth Field (Dixon *et al*., 1995) as well as the sandy volcanism. Greenland’s model (1987, 1998) has now to be integrated in this category (e.g. Surlyk, 1987, 1998; Gjelberg *et al*., London conference 2001).

C3 – The syn-depositional model assumes that all the sandy injections (sills and dykes) take place during the (very short) period of time during which the reservoir feeder is deposited on the seafloor. This model invokes only sedimentary processes. It involves a good reservoir characterisation, including its geometric relationship with the fully described injections and the paleo-geometry of the seafloor.
In this case, it will be possible to determine a post-injection palaeocompaction curve fir the host-formation (e.g. Beaudoin et al., 1987), then a palaeoporosity curve for superficial, wet sediments, from 0 to 400 metres-deep (Schneider & Parize, 1989).

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CLASTIC INJECTIONS:

PER ASCENSUM
VERSUS
PER DESCENSUM

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PER ASCENSUM

Newsom, 1903

PER DESCENSUM

Notes

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THE VOCONTIAN DOMAIN

A RARE, WELL-EXPOSED OUTCROTTING AREA FROM SHELF TO UPPER SLOPE

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From a paper submitted to Sedimentology
By G. Friès and O. Parize

LITHOSTRATIGRAPHIC SETTING

Litho-stratigraphical sub-division of the Aptian succession in the Vocontian sector is based on

(1) a number of 'key' lithological markers that are continuous and thought to be synchronous throughout the slope domain and

(2) detailed biostratigraphy, mainly based on ammonites and foraminifera (Fig.). This framework enables the identification of several breaks in deposition, in places associated with localised erosional surfaces several tens of metres deep, but elsewhere in apparent stratigraphic continuity (Fig.).

The Vocontian Aptian Blue Marls Formation

The Aptian marls belong to the so-called "Marnes Bleues" (Blue Marls) Formation (Flandrin, 1963) which appears abruptly above the thick marly calcareous succession of lower Cretaceous age. This abrupt change of lithology is usually related to a major crisis in the carbonate productivity and is not directly linked with a eustatic event (Cotillon *et al.*, 2000). The Blue Marls Formation has a maximum total thickness of 300 m and is typically composed of 50% carbonate, 30% clays and 15% quartz, with the remainder consisting of accessory minerals such as pyrite and feldspar. The clay fraction content is less than 15% chlorite, 30–40% illite, 15–25% mixed layer clays, up to 15% smectite and 20–25% kaolinite (Deconinck, 1984). The average organic content amounts to about 0.5%. The lithological homogeneity is interrupted by strong banding in which light bands a few decimetres thick alternate with dark bands, reflecting $CaCO_3$ and TOC fluctuations. This banding is an indication of a Milankovitch-type climatic cyclicity (Ferry & Rubino, 1987; Bréhéret, 1994).

The biostratigraphic framework for this study was based mainly on planktonic and benthic foraminifera (identification by G. and J. J. Bizon) In addition, ammonite collections by Friès, (1987: identification by R. Busnardo, P. Destombes and J. Sornay), together with those of Moullade (1966) and Bréhéret (1997), served to establish an ammonite biozonation similar to that for the Tethys domain (Hœdemæker *et al.*, 1993; Bogdanova & Tovbina, 1994). Recently New ammonites collections are realized with LG Bulot & JL Latil.

APTIAN Lithostratigraphic markers

Four major lithostratigraphic units (Fig.) have been identified and can widely correlated in the Aptian succession: Units B, G, K1 and K2 (Friès, 1987). The base of Unit B is marked by a change of lithology and the top by a change of colour (with intense darkening of the marls, coupled with an increase in the TOC content; (Fig.). Unit G starts above the change of colour and is capped by the 'Nolan beds', a distinctive carbonate horizon. It is followed by Unit K1 which is capped by the 'Fromaget beds' carbonate horizon (Fig.). The uppermost Unit K2 is bounded at the top by another major change of colour (intense darkening of the marls).

Three main horizon of calcareous beds interbeded with marls have been identified and traced throughout the Vocontian domain (Friès, 1987; Bréhéret, 1997): the White beds (Fig.), the Nolan beds and the Fromaget beds (Fig.). The first unit is 1 metre thick and the following two are 10's m thick. The Fromaget beds provide a high-resolution datum for the analysis of $\Sigma 2$ to $\Sigma 7$ slumps (Friès, 1987) and to the locally injected sandstone sills and dykes found in this interval (Paternoster, 1983; Beaudoin *et al.*, 1983; Parize, 1988).

There are also four distinctive layers enriched in organic matter that have total organic content (TOC) values of 2–5%. These layers, a few centimetres to several tens of centimetres thick, form a number of marker beds (Fig.): the Goguel, Fallot, Jacob and Kilian levels (Bréhéret, 1997). The Goguel level is the time equivalent of the Lower Aptian Oceanic Anoxic sub-Event OAE1a (Bralower *et al.*, 1994).

Several yellowish smectitic layers a few centimetres thick (smectite content between 50 and 100%) have been discovered since 1995 (Beaudoin & Friès, oral communication) and described: examples are the Van Gogh (Fig.), Cézanne and Matisse levels (Dauphin, 1997; Dauphin, *in prep.*). They are regional in their extent, are interpreted as alteration products of volcanic ash beds and may be widely correlated across Western Europe (Spain, West Germany, Great Britain, etc: Dauphin, in prep.).

The gravity-driven deposits studied here sit mainly within Unit G (A to J slumps; P2 and T1a & b turbidite bodies). However, two massive slumps occur in both Units B (α and γ slumps) and K1 ($\Sigma 1$ and $\Sigma 2$ slumps); a massive sand body sits at the base of Unit K2 (T3) (Fig.). The maximum thickness of these gravity-driven deposits and the depth of erosion associated with them are shown schematically in (Fig.).

ALBIAN Lithostratigraphic markers

APTIAN-ALBIAN SEDIMENT GRAVITY FLOW DEPOSITS

The AptianAlbian marly pelagic succession is punctuated in the Vocontian area by various gravity flow deposits. These mass wasting or mass movement deposits include all resedimented material emplaced by the force of gravity ranging from slides to turbidites (*e.g.* Nardin *et al*., 1979; Stow, 1992) *sensu* 'subaqueous gravity deposits' as used by Dott (1963). In some sections, such as those at Arnayon (Fig.) and La Charce (*id.*), they correspond to more than 90% of the Aptian succession, whereas for others, such as Serres-Chaitieu (*ibid.*) this proportion is less than 10%. The importance of resedimentad deposist decrease during Albian and increase suddlely during Vraconian

The various gravity-controlled deposits are grouped into two categories: sandy gravity flow deposits and mass movement complexes or mass transport complexes (Weimer, 1990, but used here without any sequence stratigraphic connotation), the latter including slumps and debrites. Clastic injectites are linked to massive sand deposits that will detail now.

These massive sands are one type of sandy sediment gravity flow deposits in vocontian area. We include in this group all turbiditic sandbodies (Middleton & Hampton, 1976; Lowe, 1979; Postma, 1986; Mutti, 1992). We have defined two associations on the basis of the relative importance of the massive sandstone facies (*e.g.* Aboussouan, 1963; Rubino, 1989) versus differentiated turbidites *sensu* Bouma or 'classical' turbidites, (Bouma, 1962: discussion *in* Walker, 1992 and Mutti, 1992). The average composition of the sandstones in both associations is 60–70% quartz grains, with 20–30% detrital glauconite grains, and the remainder consisting of siliceous and carbonate shell debris, some feldspar, detrital mica and infrequent fragments of tourmaline and zircon. The grain size is typically 100–300 µm, the clay content of the sandstones being negligible (Aboussouan, 1963; Rubino, 1981; Parize, 1988).

Four massive sandstone bodies have been recognised within the Aptian to Lower Albian succession: two sit beneath and considered part of the T1a and T1b turbidite bundles in Unit G. The T3 body occurs in Unit K2 whereas the T4 body sits in first Lower Albian unit (Fig.). These massive sandstones are invariably thicker than the previously described turbidite bundles and internal sedimentary structures are characteristically scarce:
*** in the albian

Description: Massive sandstones form 90% to as much as 99% of the total thickness of the sandbodies. The sand texture is fine to medium-grained, with occasional scattered quartz granules a few millimetres in size. Where sedimentary structures such as parallel laminations and current ripples occur, they are confined to the uppermost part of the bodies. They are sometimes distorted by intense liquefaction or dewatering features. Marly calcareous pebbles a few centimetres across represent rip-ups from the pelagic substrate and form the *nuclei* of nodular cements that are found at the top of the T1a body and in the massive part of T3.

Individual massive sand beds are 5-30 m thick (typically 10 m) and exhibit sharp and flat to strongly scoured bases from their proximal to most distal part (Imbert, Parize & Rubino, unpubl. data). They stack vertically to form massive sandstone units, often without visible signs of erosion and amalgamation; these units reach a thickness of 40 m in the Upper Aptian (T3: (Fig.); Parize, 1988; Rubino, 1989). The superposition of two successive massive sand beds (*e.g.* T1a) is sometimes accompanied by spectacular loading at the amalgamation surface (*e.g*. Hiscott & Middleton, 1979). Linkage of sandstones beds occurs by either the development of flame structures, or by basal metre deep, concave–up scours (T3). In the latter case, the junction between two separate beds is no longer discernible as the thin bed capping interval with sedimentary structures (if one existed) is eroded (*e.g.* Walker, 1966).

The massive sandstone beds fill erosive channels which are several hundreds of metres wide and tens of metres deep (Fig.). The erosion surfaces have associated small-scale scours which cut into the marly substrate. They are metre-deep and are most frequently found at the transition from the bottom to the sides of the channel-form erosion surfaces. In each channel fill, the lower massive megabeds are covered by ‘classical’ metre thick turbidites (*e.g*. upper part of T1a and T1b).

Metre-thick packages of decimetre-thick sandy beds occur outside but immediately adjacent to the T3 channel (Parize, 1988). These deposits, which form a hundred metre-wide belt, preserve no internal structures and they have wavy bed bases and tops.

The presence of extensive sandstone sills and dykes which have been injected laterally from the channelised erosion surfaces into the banks is the most a distinctive feature of the massive sandstones units (Parize, 1988; Parize *et al*., 1999). The most spectacular injections are found in the T3 deposition system (Fig.); Paternoster, 1983; Beaudoin *et al*., 1983; Friès *et al*., 1984; Parize, 1988). The sills, which may be several metres thick, are closely associated with the dykes (Fig.). This association exhibits a characteristic staircase geometry that distinguish sills from true beds (Fig.), all the more so because the sandstone sills are injected into the mud-prone channel banks (Fig.) and extend for several kilometres laterally.

Interpretation

Well differentied from 'classical' turbidites, these Vocontian massive sandstone deposits have been previously described as submarine sandy avalanche (Aboussouan, 1963). They resemble Facies B of other workers (Mutti & Ricci Lucchi, 1972; Pickering *et al*., 1986; Walker, 1992), and units described as fluxoturbidites (Slaczka & Thompson, 1981) or sandy debris flows deposits (Shanmugam *et al*., 1994; Shanmugam & Moiola, 1995; Shanmugam, 1996; Stow & Johansson, 2000).

A cursory examination of the outcrops of the T3 package might give the impression that the massive sandstone bodies comprise sediment blocks torn from the substrate and transported without any great disruption, resembling fluxoturbidites (Slaczka & Thompson, 1981; Beaudoin *et al*., 1983; Friès *et al.,* 1984) or sandy debris flow deposits (Hiscott & Middleton, 1979; Shanmugam *et al*., 1994; Shanmugam & Moiola, 1995; Shanmugam, 1996; Stow & Johansson, 2000). In fact the 'blocks' correspond to the channel banks injected by sills and dykes. The Vocontian massive sandstone units are interpreted as high-density turbidity current deposits (*sensu* Lowe, 1982) characterised by sharp erosive bases and tops, a homogeneous texture (some fine grading may nevertheless be present) and a lack of sedimentary structure (Parize *et al*., 1999). The sandy turdidites fill slope incisions reaching up to 40 metres deep with vertical/steep flanks and flat floors. Undercuts on the incision margins a few m's deep are locally observed. The channels have widths ranging from 200 to 1000 metres. The textural uniformity of the sand is seen both longitudinally and transversally across the system and is thought to reflect a vigorous winnowing process on the shallow water platform which led to a particularly homogeneous, fines-deficient sand supply (Imbert, Parize & Rubino, unpubl. data).

The massive sandstone bodies can be followed for up to 50 km downslope and are channelised right to the most distal sections where concave–up scours and undercuts are well developed. Mutti (1992) notes that concave–up scours are common close to the transition between channel-fill and channel-lobe deposits and that they are the product of intense turbulence associated with the transformation of fully turbulent turbidity current (Hiscott & Middleton, 1979). The massive sandstone show little lateral downslope change in structure, unlike mixed sand-mud systems, and the only perceptible indication of distality is the predominance of sand injections in more distal areas. Fluid-expulsion structures (dish structures) have been described only in the sands of the transition zone between the channel filling and the injections, *i.e*. close to the channel margins.

The thin-bedded sandstones associated with the T3 body could be indicative of by-passing, but they are interpreted as limited channel overbank deposits. Two factors may account for the subdued thickness/extent of overbank units in the succession. Firstly, the channels could have been too deep for the current to spill out, even when swollen by incorporation of water. Secondly, the high sand/clay ratio of the flows may have meant that the flows were relatively thin (Crémer, 1983; Stow *et al.*, 1983; Laval, 1988).

The rates of compaction measured directly from the geometry of the injections (*e.g.* ptygmatic folds) demonstrate that the injection took place under only a few tens of metres of sediment (Schneider & Parize, 1989). The sand injection was thus an early phenomenon. All injections have been observed below the stratigraphic level of the top of the massive sand body to which they are linked (*i.e.* the palaeoseafloor) and are only found laterally to the channels.

Wider spatial organisation of the sediment gravity flow deposits

The mapped geometry of the Aptian mass transport complexes (Fig.) shows that each covers several hundred square kilometres. Their thickness decreases downslope; however, local thickness variations are associated with the position of syn-sedimentary faults (Friès *et al*., 1985).

Starting from the transition between the shelf and the slope domain, the main sandy gravity flow deposits are traceable for 15 km (T2), 40 km (massive facies of T1 and T3) and 80 km (P1, P2 and upper part of the T1 deposits) downslope. These sandbodies are channelised, even in the most distal slope succession exposed (Rubino, 1989). The various thin-bedded turbidite sets, P1, P2 and the upper part of the T1 (Rubino, 1989; Friès & Rubino, 1990) correspond to highly efficient flows, while the massive sandstones correspond to deposits of rather inefficient flows *sensu* Mutti (1979).

The massive sandstones (T3 and T4 and the lower part of T1) backfill several hundred metre wide erosive channels using the channel classification established by Normark (1970) and Mutti & Normark (1987, 1991). The classical turbidite bundles (T2 and the upper part of the T1) occupy channels characterised by mixed or divergent styles of filling *sensu* Mutti & Normark (1987, 1991) in 3 to 5 km wide topographic lows. The thin P1 and P2 bodies have the greatest mapped extent, both transversely and longitudinally but are superimposed on the axes of earlier T1, T2 and T3 channel fills.

Pelagic and gravity-driven sedimentation on the Vocontian slope margin was predominantly marly and marly calcareous in character: sandy facies account for only 10% of the succession. The lithological bias towards finer grained deposits is even more striking when the volume of individual systems is taken into account. For example, Σ2 slump represents 60 km^3 of reworked (and compacted) sediment, whereas the associated T3 sandy system represents only 0.5 km^3: Tertiary turbidite systems of the North Sea which were of comparable length have on average 50 times more sand (Reynolds, 1994). According to recent classification schemes (Reading & Richard, 1994; Richard *et al*., 1998**)** and the hierarchy proposed for depositional turbidite units (Mutti & Normark, 1987, 1991)**,** the Aptian gravity deposits of the Vocontian palaeo-margin belong to a mud-rich complex with episodic access to sand allowing short-lived sand-rich systems to develop (T1, T2, T3). This reflects an important switching in source, with the sandy sediment gravity flow deposits produced by erosion of a shallow water shelf on which sand was locally stored and 'polished', and mass transport complexes which were delivered by failure of the upper slope domain, or even the outer platform when this was mainly mud-prone: it is easier and quicker to obtain a massive sand if the source is already sandy and sorted (Imbert, Parize and Rubino, unpubl. data; Parize *et al*., 1999).

The Vocontian domain exhibits a mode of organisation with gravity-driven sedimentation dominated by argillaceous sandstone and marly mass transport complexes (Rubino, 1989; Friès & Rubino, 1990). The subordonate sandy units consist of confined channelising deposits which were mainly erosive from the proximal to the most distal parts (Parize, 1988; Rubino, 1989).This organisation may not be applicable to the large present-day turbiditic systems (Zaire, Amazon, Nile deep sea fans, etc.); it does, however, incorporate very similar constituent architectural elements:

(i) large-scale erosional features,

(ii) channelised or confined deposits,

(iii) slumps, debrites, massive sands, Bouma sequences and sand injections and

(iv) pelagic and hemipelagic deposits and differents types of back-shales.

Sequence stratigraphy of Aptian albian serie

The slumps, sandy turbidite deposits, condensed levels, changes of colour, major erosion surfaces and biostratigraphic hiati have been used to erect a sequence stratigraphical sub-divison which allows the slope and platform segments of the Vocontian margin to be correlated (Rubino, 1983, 1989; Friès & Rubino, 1990).

Four depositional sequences (*sensu* Vail *et al.,* 1977, 1991; Van Wagoner *et al.,* 1988) are identified: Ap1 to Ap4 (Fig.). The main criteria (Rubino, 1989; Friès & Rubino, 1990) used to define the slope depositional sequences and their component system tracts (*sensu* Brown & Fischer, 1977; Posamentier *et al*., 1988) are: (i) the large-scale erosional regional surfaces define the sequence boundaries, (ii) the various limestone bundles (White beds, Nolan beds and Fromaget beds) are interpreted (Rubino, 1989) as lowstand prograding wedges (discussion *in* Kolla, 1993) and (iii) the condensed intervals (picked out by colour changes and overlying black shale levels) which extend onto the shelf (Friès, 1987; Rubino, 1989; Bréhéret, 1997) represent maximum flooding surfaces.

Depositional sequence Ap 1

The identification of the Ap1 sequence boundary in the Uppermost Barremian-Lower Aptian (Ferry & Rubino, 1989) is problematical as it coincides with a drowning event, a change in carbonate productivity (Cotillon *et al*., 2000) and a major erosional mass transport complex (Ferry, 1976; Friès, 1987; Bréhéret, 1997); in combination these events led to a dramatic change in lithology from pure limestone to almost pure marl deposits. The Goguel level and White beds are geographically restricted to the pelagic domain (Fig.) and belong to the lowstand system tract of the Ap1 sequence (Rubino, 1983, 1989; Friès & Rubino, 1990). The interval between the White beds and the next colour change corresponds to the transgressive system tract. The colour change represents the maximum flooding surface and it has been traced on the coeval platform (Rubino, 1989; Friès & Rubino, 1990). The Ap1 highstand system tract is well developed: during this tract and in the Crest area (Fig.), the outer shelf has prograded basinward by up to 10 kilometres.
The first middle Aptian slump set (A to D slumps, (Fig.) records the destabilisation by slope failure of this prograding shelf. This slump set can be interpreted as a megaslide system (Galloway, 1998). The slumps are interbedded in the Fallot level (Bréhéret, 1997; Dauphin, *in prep.*) which presents enrichment in organic matter of type III (continental origin). The A slump sits locally on the White beds (Sig and LM sections, (Fig.) with the total or partial absence of the *cabri* and *ferreolensis* zones and local erosion of 40 m of the pelagic deposits (Fig.).

Vocontian Albian « Blue Marls »

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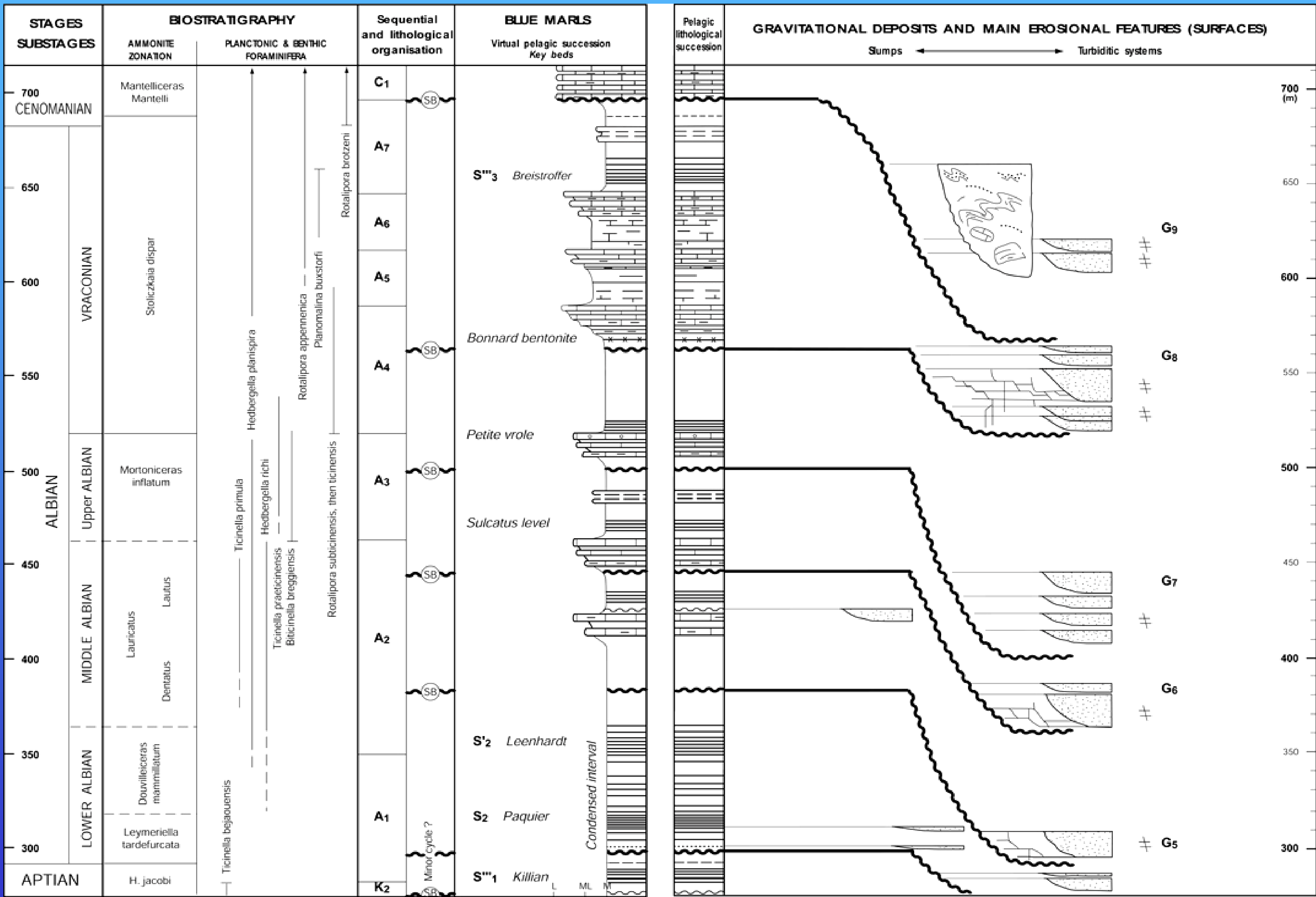

Vocontian Aptian « Blue Marls »

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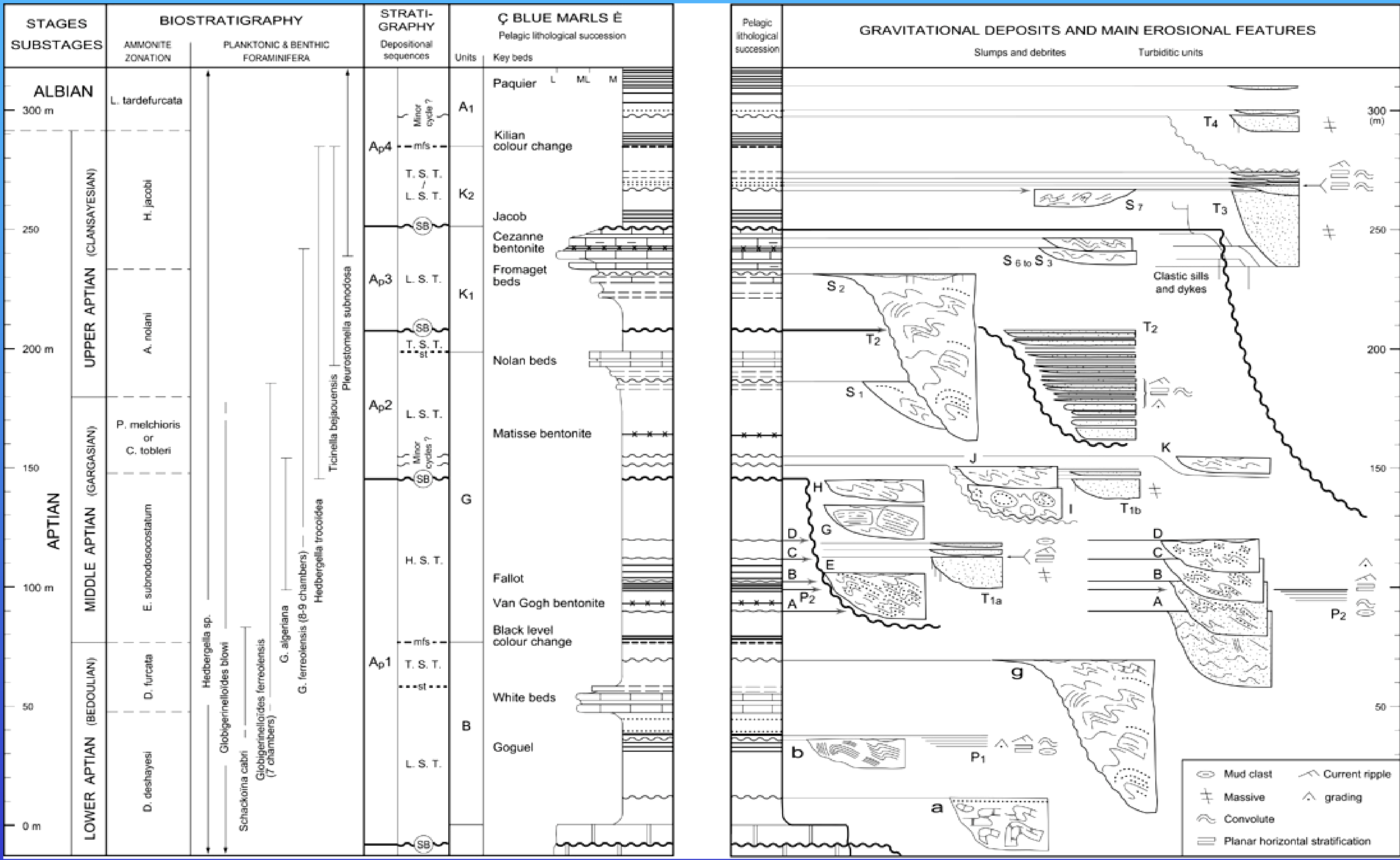

Depositional sequence Ap 2

An erosion surface at the base of the second slump set (E to K) represents the first well defined Aptian slope sequence boundary. The base of this sequence is variably characterised by reworked silty and marly thin-bedded turbidites assumed to be by-pass deposits, E slump and major clastic influx (T1a and b channels). A number of minor sequences are defined within this central unit (see below). The Nolan beds represent the lowstand prograding wedge of the Ap2 sequence; the □1 slump records a slope failure during this progadational phase. The Nolan beds are capped by a thin marly transgressive system tract (Fig.).

The sediment gravity flow deposits of the Ap2 lowstand system tract (E to K slumps; turbiditic bodies T1a and T1b, (Fig.) appear to be organised into three higher frequency sequences. The base of the first of these higher frequency sequences (which includes E to H slumps, and the T1a channel fill) is characterised by considerable erosion (upper *algeriana* zone, (Fig.); up to 60 metres deep) which cuts deep into the underlying hemipelagic marls. The second sequence (I and J slumps and the turbiditic system T1b) is also bounded by a basal erosion surface. The upper higher frequency sequence is only represent by the K slump. The slumps/debrites (E, I and K) linked with the erosional surfaces can be related to mass transport complexes *sensu* Weimer (1990).

These three high-order sequences have a similar spatial distribution with vertical superposition of the axes of remobilised units and a progressive backstepping pattern of deposition. Their organisation indicates an upward decline in the importance of basal erosion and the number of slumps and the volume of reworked sand. They can be related to turbidite stages *sensu* Mutti & Normark (1987). The two turbidite sand bodies T1a and T1b represent channel-fills developed during the backstepping infill (Rubino, 1989) as theoretically suggested by Mutti (1985) and Mutti & Normark (1987); they record the first influx of sand in the form of massive sandbodies on the Vocontian slope.

Depositional sequence Ap 3

An erosional surface of regional extent is locally associated with the T2 sandstone package (Fig.) and defines the lower boundary of the Ap3 depositional sequence. This unconformity is associated with the local development of diagenetic concretions in the underlying marls (Friès, 1987; Bréhéret, 1997). The lowstand prograding wedge is well developed with the 10-30 m thick bundle of the Fromaget beds. The progradation of this formation reaches 10 kilometres basinwards of the Marsanne fault to the Mornans fault, in the Crest area (Fig.): the Σ2 to Σ6 slumps record important slope instability during this progradational phase. The

$\Sigma 2$ slump is the largest sediment gravity flow deposit in the Aptian slope succession (see below). Because this large slump which can be associated to one event, with a characteristic internal organization, redefine the regional slope morphology and the shelf margin position and it leave a major erosional surface, it can be related at the same time to a megaslide system *sensu* Galloway (1998) and a mass transport complex *sensu* Weimer (1990).
The depositional sequence Ap3 is truncated on top of the Fromaget beds by another regional erosional surface. Therefore, the Ap3 transgressive and highstand system tracts have not been preserved in the Vocontian area.

Depositional sequence Ap 4

The base of sequence Ap4 is marked by a tectonically enhanced unconformity (Friès, 1987) which locally erodes down to the marly highstand deposits of the sequence Ap1 (Fig.). This erosional surface is covered by the Jacob Level (Fig.) and then by the T3 erosive massive sandstone unit. Near the top of the Aptian stage a change of color below the Kilian black shale bed represents a second maximum flooding surface within the pelagic succession. This lower part of sequence Ap4 represents the lowstand and/or the transgressive system tracts. The thin marly high stand system tract already belongs to the Lowermost Albian.

Depositional sequence Al 1

Depositional sequence Al 2

Depositional sequence Al 3

Depositional sequence Al 4

Depositional sequence Al 5

Summary

Four depositional sequences are defined in the Aptian slope succession. Sequences Ap1 to Ap3 are increasingly poorly preserved due to an increase in the depth of basal erosion. By analogy with present-day data (Mulder & Cochonat, 1996; Piper *et al.*, 1999), this slope erosion is likely to have been retrogressive and to have triggered successive failure and sliding. An idealised Vocontian slope sequence comprises: (i) a prominent basal unconformity, which may be covered by a turbiditic package (*e.g*. T2), a slump (*e.g*. E slump) or a black shale bed (*e.g*. Jacob Level), (ii) a limy lowstand prograding wedge reflecting slope extension, variably reworked by slope failure, (iii) a thin marly transgressive interval, (iv) a maximum flooding surface and (v) a marl-dominated highstand system tract where slope instability is recorded by numerous gravity flow deposits.

The various organic rich bundles recognised in the Aptian slope succession may have different stratigraphic implications (*e.g*. Fiet, 2000). Some may be associated with maximum flooding surfaces (Black level, Kilian level), but others overlie major erosional surfaces (Jacob level) and therefore they may have the same significance as the Quaternary unifites from the Mediterranean Sea (Feldhausen *et al*., 1981).

The Aptian Vocontian sequence stratigraphy model (Fig.) improves the previous charts *e.g*. Cretaceous 1988 chart (version 3.1A) by B.U. Haq, J. Hardenbol, P.R. Vail, L.E. Stover, E.C. Wrignt and R. Jan du Chêne *in* Haq *et al.* (1988). The four Aptian Vocontian sequences are certainly third-order sequences related to Hardenbol *et al*. chart (1998).

Sequence stratigraphic organisation of unstable slopes

The studied Vocontian slope sections reveal an alternation of argillaceous sandstone and marly mass transport complexes, pelagic slope sediments and subordinate sandy units in the form of channelised deposits which were erosive from the platform-slope break to the lower slope.

The sequence stratigraphic model, more frequently called “Exxon model”, for the evolution of passive margin slopes (*e.g*. Vail *et al*., 1991) envisages that:

(i) large sandy turbidite systems overlie depositional sequence boundaries, at least on the lower slope and basin floor,

(ii) debrites and slumps characterise the lowstand prograding wedge (see the discussion *in* Kolla, 1993 and Bertram *et al*., 1996),

The Vocontian slope sequences demonstrate that on the slope, the sequence boundary may be associated with either:

the base of large muddy debris flows,

the base of sandy turbidite packages or even a black shale horizon (*e.g*. Bertram *et al*., 1996 and Fiet, 1999).

Thus, at outcrop scale, the sequence boundary is hard to pick without a detailed understanding of the shelf-slope stratigraphy constrained by high resolution biostratigraphy whereas at seismic scale it is necessary to well correlate large erosional features with sedimentological data from cores. The development of large instability features on the slope reflect rapid progradation of the shelf during the highstand system tract (A to D slumps) and delta build out to the shelf break during the lowstand prograding wedge (Σ1 and Σ2 slumps).

The sequence stratigraphic model envisages that turbidites are directly fed during the sea level fall through large rivers incising the shelf, whereas the influence of contour currents and shelf drifting is ignored (see the discussion *in* Mutti, 1992) or by unstabilities of delta front deposits implying genetic continuity from slides to 'fluidal flow deposits', taking in 'mass flow deposits' (*e.g.* Dott, 1963; Nardin *et al*., 1979), as notably illustrated by the 1929 Grand Banks Event (Heezen & Ewing, 1952; Kuenen, 1952; Normark & Piper, 1991; Piper *et al*., 1999). On the Aptian Vocontian palaeomargin, the sequence boundary in not linked to valley incision on the shelf and the vocontian turbidites were derived from shelf sand stores rather than shelf edge deltas. The clean texture of the sands meant that these produced low efficiency flows which were unable to bypass the long slope segment; consequently, a significant feature of this slope system is that sands (albeit subordinate to slumps and debris flows) were deposited on the slope, and locally confined even to the upper slope. Although the Vocontian slope system can be followed for over 80 km down slope from the shelf edge, material removed during periods of major slope erosion (healed by the gravity driven deposits described here) cannot be found in the most distal outcrops. This material must have fed a rise/deep sea fans on the edge of Valais ocean but this part of the system has not been preserved. Thus, the preserved slope deposits record the retrogressive infilling of large erosive valleys (50 m deep, 5 to 10 km wide, at least 100 km long) cut into the slope but the products of slope cutting are lost.

The Aptian Vocontian slope system represents an alternative to the 'classic' EXXON mud-rich delta-fed systems model (Van Wagoner *et al*., 1988; Vail & Sangree, 1988). Key elements of the Vocontian reconstruction are

(1) an emphasis on lowstand slope erosion and complex slope morphology controlled by contemporary tectonism and salt diapirism,

(2) slope deposition in confined erosional and structurally controlled conduits rather than the buildout of slope fans/channel-levee complexes

(3) a dominance of large volume muddy slump and transitional debris flow deposits, with suboridate sandy turbidites, including significant massive sandstone facies

(4) intermittent sand supply by collapse of shelf sand stores with significant sand deposition on the upper slope and heavily sand-injected downslope fringes to the sandbodies and

(5) minimal downslope change in turbidite facies character and little evidence for significant flow transformation.

THE VOCONTIAN SLOPE: AN EXAMPLE OF TOPOGRAPHIC CONTROL OF GRAVITY SEDIMENTATION

There is good evidence that synsedimentary tectonics influenced slope morphology and lithology distribution on the Vocontian margin. In the Rosans area, the Aptian turbiditic system diverged around the emerging Vaucluse anticline, onlapping onto the flanks of the anticline (Rubino, 1983, 1989; Parize, 1988). On the Provence platform, the Albian sand channels associated with the Banon fault network (Fig.) followed a grabenal structure traditionally assigned an Oligocene age (Debrand-Passard *et al*., 1984). However, this rift is associated with deep sinistral strike-slip faulting which was active since the basal Aptian (Joseph *et al*., 1987), with the depression maintained by differential compaction throughout the Albian (Joseph *et al*., 1987).

The slump units also exhibit significant transverse variations in thickness (by more 25%) which were controlled by the position of syndepositional anticlines and synclines (Rubino, 1983, 1989; Friès, 1987; Parize, 1988). The maximum depth of erosion beneath the slumps is found in synclinal zones and the best stratigraphical continuity of the pelagic deposits is found on the flanks of the anticlines, a relationship inferred in the 19th Century (Gras, 1835). Isopach maps for each of the main gravity flow bodies established from several tens of sections covering the sector (Friès, 1987) and taking account of late Alpine tectonics (Goguel, 1947; François, 1981; Friès, 1987) demonstrate an original topographic control on stacking of the slope remobilised deposits.

Palaeogeographic reconstruction and evolution of the palaeomargin

There were four main pathways routing sediment downslope (Fig.):

(1) the palaeovalleys issuing from the Dauphiné platform;

(2) the large feeder system rooted in the corner between the Dauphiné platform and the Vivarais platform;

(3) the palaeovalleys issuing from the Vivarais platform: **Rosans turbiditic system**;

(4) palaeovalleys issuing from the Provence platform: **Bevons turbiditic system**.

The large Céüse complex in the north-western 'corner' of the Vocontian margin (Fig.) was active throughout the Aptian. The restored palaeotopography here corresponds (Fig.) to a cluster of bathymetric lows interpreted as submarine palaeovalleys and separated from each other by strings of elongated highs corresponding to the crests of fault blocks (Joseph *et al*., 1989) and/or salt diapers or ridges (Martinod, 1988; Parize, 1988). The highs are cut by 'cols'

Paleomorphology of Vocontian basin

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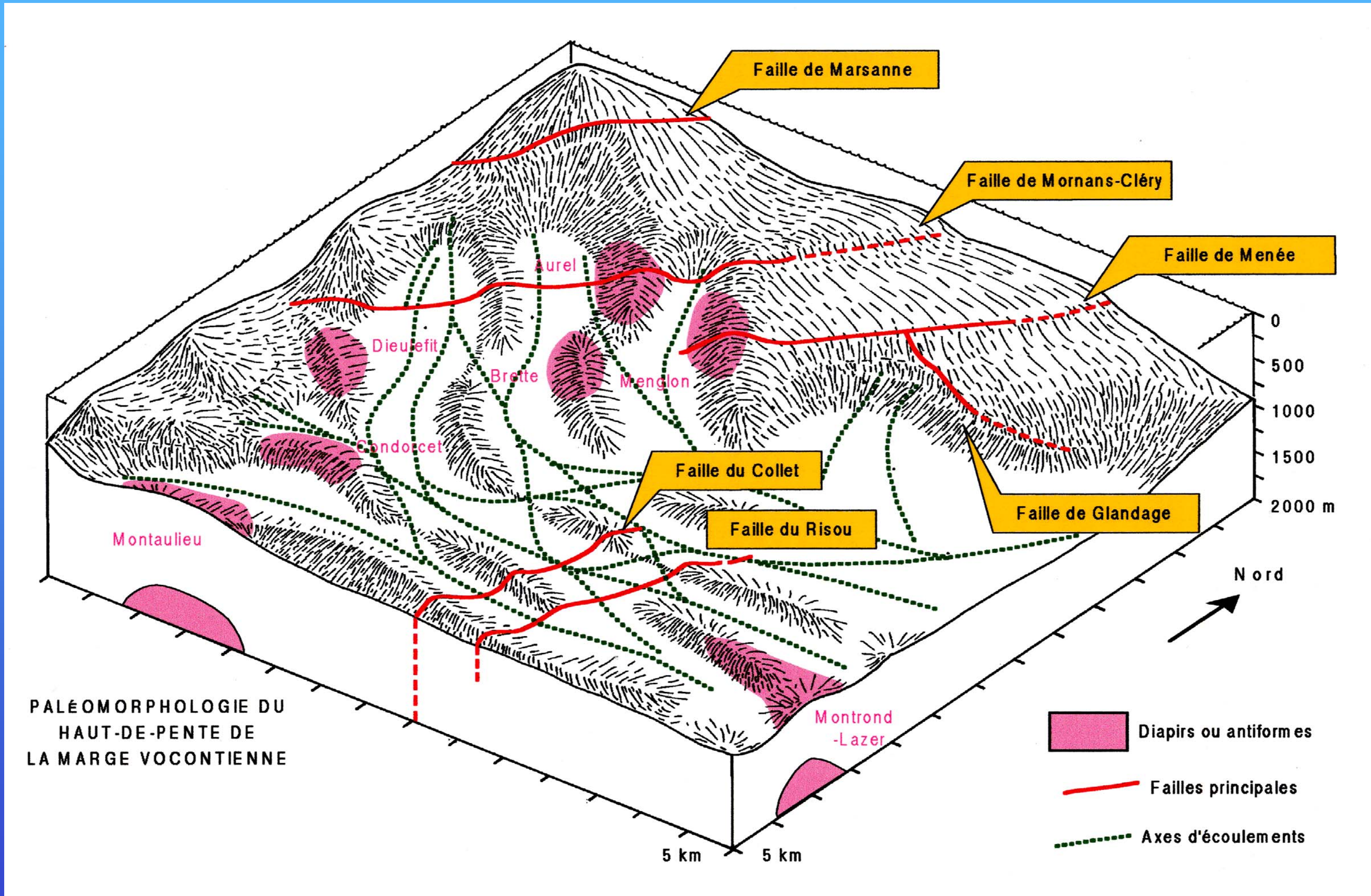

that allowed the passage of some gravity flows from one palaeovalley to another. The bathymetric difference between the top of the highs and the bottom of the lows or channels would be of the order of several tens of metres, because the slumps are able to bury the entire area, whereas the sands are always strictly channelled.

The fence diagram (Céüse 3D Block) reconstructs the overall geometry of the slope complex flattened at the White Beds level. The three-dimensional reconstructions visualise the geometry of the margin, the part played by major penecontemporaneous faults, the palaeo-highs between the different submarine valleys and the extent of the sedimentary bodies. A palaeo-water depth of 1000 metres in the lowest currently known part of the slope domain in the Gap region (Fig.) is inferred (see earlier discussion).

The Vocontian slope was characterised by an abundance of submarine gravity-driven deposits during the Aptian, mainly slumps which transformed to debris flows. Downslope confinement in submarine valleys several kilometres wide and at least several tens of kilometres long explains both the vertical stacking of the slumps and turbidite sandbodies, and the pronounced lenticularity of the deposits in cross-section. The sea floor appears to have been most rugose in the Middle Aptian: this was the time when the slumps were most numerous, with repeated episodes of deep erosion. The turbiditic systems attained there greatest longitudinal dimension at this time, reaching 80 km from the slope break.

In the north-western part of the Vocontian slope, the various submarine valleys in the Aptian appear to have been long-standing features, functioning as conduits for more than 40 million years from the Berriasian to the Cenomanian (Friès & Beaudoin, 1985a; Joseph *et al*., 1989). This longevity reflects maintenance of bathymetry by penecontemporaneous tectonic activity, locally accentuated by salt tectonics. The local distribution of the sedimentary bodies was also controlled locally by differential compaction (Friès *et al*., 1984).

Role of tectonics

The Aptian in SE France was characterised by the drowning of the Urgonian platforms, the onset of the first siliciclastic sedimentation since the Triassic, and increased morphological complexity on the platform margin slopes. These events would appear to be a consequence of the Austrian compressive phase which affected much of Western Europe (Bénard *et al*., 1985; Ziegler, 1990). This compressive phase coincided with the start of the oceanic expansion in the Gulf of Biscay and the rotation of the Iberian plate. In SE France, compressive events are recorded by Cenomanian and later folding and the development of east-west trending reverse faults (Friès, 1987). The more complex slope topography in the Aptian may be linked to the increased growth rate of the structures relative to sedimentation

Céüse stratigraphic 3D block

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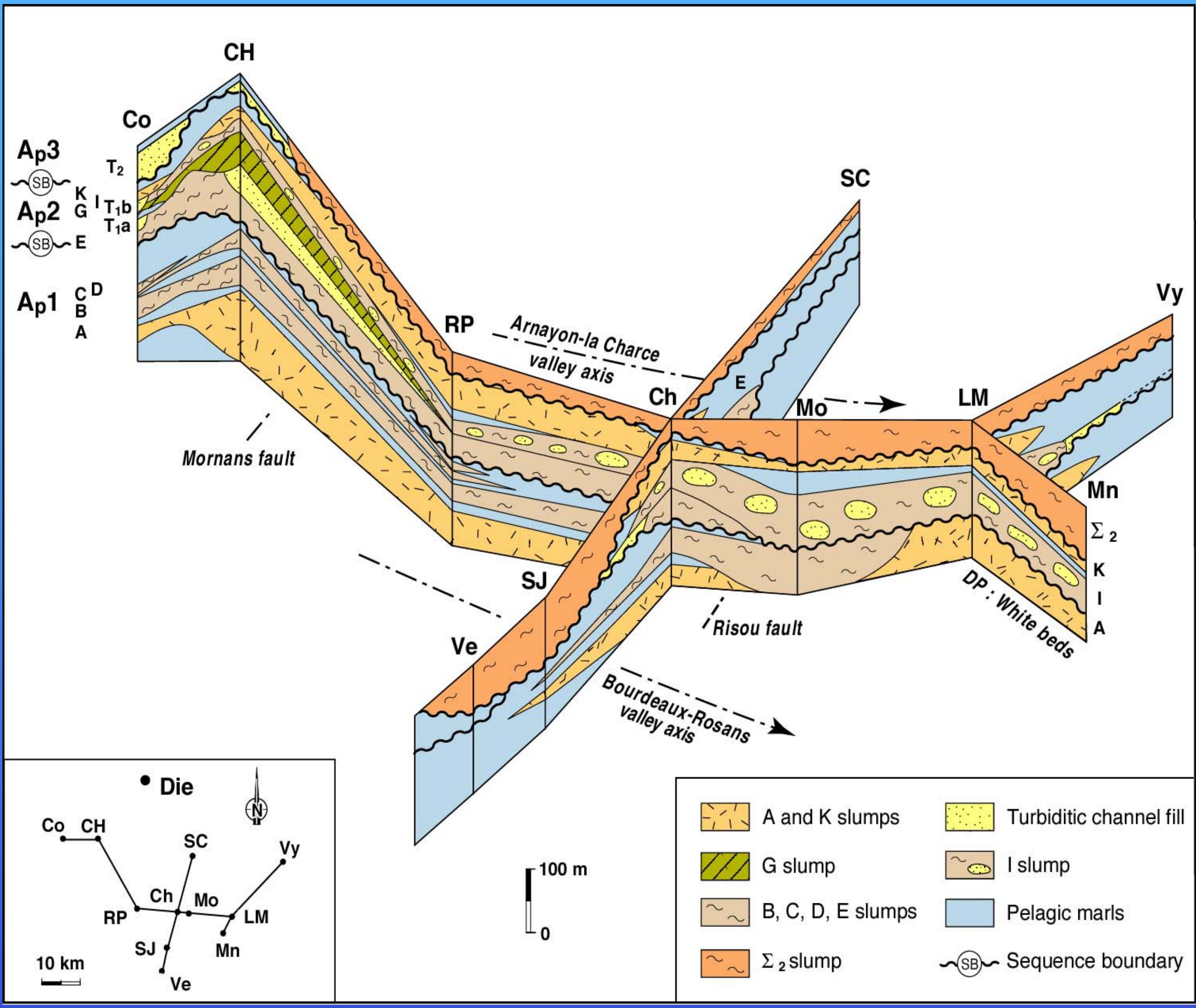

rates (*e.g.* Apps *et al*., 1994), and this in turn can be related to tectonic inversion which began in the area in the early Aptian (based on microtectonic data of Bouchet, 1985 and Cabrol, 1985).

The large Cévennes, Nîmes and Durance faults were major features throughout the Mesozoic and the Cenozoic (Debrand-Passard et al., 1984). A role for these faults on sediment distribution has been demonstrated from the Lias (Lemoine *et al*., 1986; Stampfli *et al*., 1998) through to the Holocene. All these faults (*e.g.* the Cévennes, Nîmes, Durance faults; (Fig.) have an important penecontemporaneous activity both on the platforms (Beaudoin *et al*., 1986; Montenat *et al*., 1986; Joseph *et al*., 1987) and in the slope domain (Friès *et al*., 1985; Friès, 1987) during the Aptian The normal component of this faulting is generally antithetic to the slope (Friès *et al*., 1985).

The $\alpha, \gamma, \Sigma 1$ and $\Sigma 2$ slumps that rework carbonate bundles are not confined to the north-western part of the slope domain and do not participate in the filling of any sizeable pre-existing erosional morphology. The α slump is also identified in the south-eastern part of the slope domain (Graciansky *et al*., 1972; Friès, 1987) in the footwall of a fault. The γ slump that reworks the White beds has been identified in a large part of the pelagic domain, including the Sisteron sector (Fig.) and the south-eastern margin, where it seals a local fault. The $\Sigma 2$ slump is also not restricted to the north-western sector and has been identified in the Digne thrust unit (Fig.). These slumps appear to be major regional events; penecontemporaneous faults are always to be found at their upslope limit. Instability arising from sedimentary loading on the slope led to failure that was probably triggered by an event outside the basin (possibly seismic activity: Friès & Beaudoin, 1989), as in the case of the megabreccias on the Nicaraguan Rise (Hine *et al*., 1994).

Role of salt tectonics

The control on lower Cretaceous sedimentation by Triassic salt diapirs, perhaps accompanied by shale diapirism in the very thick shales of the Upper Jurassic (the so-called 'Terres Noires'), can be demonstrated in some localities in the South East Basin (Desmaison *et al*., 1988; Parize, 1988): the Condorcet and Propiac diapirs also exhibit end-Jurassic activity (Desmaison *et al*., 1988). Analysis of seismic profiles in the western part of the zone (Martinod, 1988) demonstrates the existence of a diapir below Dieulefit and identifies Aptian activity contemporaneous with that of the neighbouring Montaulieu and Condorcet diapirs (Fig.).

In the Aptian, the course of the T3 turbiditic system was in part controlled by this diapirism. The platform sand which was remobilised to deeper water at this time was located between the two active diapirs (Dieulefit and Montaulieu diapirs). The position of the sand filled channel on the slope was also linked to the growth of structures with diapirs at their cores, such as the Condorcet antiform (T3 sands skirt it to the north of this structure) or the Vaucluse anticline (where the turbiditic channel containing the T3 sandbody splits into two branches). The Vocontian slope morphology was therefore partly contolled by salt mobility which affected the organisation and distribution of the deposits, especially the turbiditic channels, in a comparable way to examples from the Brazilian margin (Brown & Fischer, 1977), the North Sea (Hodgson *et al*., 1992) and the Gulf of Mexico (Bouma, 1982; Apps *et al*., 1994).

Role of differential compaction

Within the same slope valley, differential compaction was responsible for the lateral offset of successive gravity-driven deposits (Brown, 1975; Pinoteau, 1986). This may be observed at two levels. Firstly, very thick bodies, such as Σ2 slump, are characterised by a convex-up upper surface reflecting subdued compaction associated with the accumulation of the largest rafts and blocks in the axial region of the slump; subsequent deposits were deflected around and accumulated on the flanks of the earlier slump bodies. This lateral offset is evident in the position of the Σ3 slump which is located on the southern flank of Σ2 slump (Friès, 1987). Secondly, on the scale of the different sequences, slumps of Lower Aptian age (α, β and γ slumps of Unit B) and Upper Aptian age (Σ2 to Σ6 slumps, Unit K1) are found in the Rosans axis, whereas the bulk of the Gargasian bodies (A to K slumps) are restricted geographically to another axis at the foot of the Dauphiné platform (Fig.).

Notes

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Notes

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DAY 1:
ROSANS AREA

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Day 1 Stops

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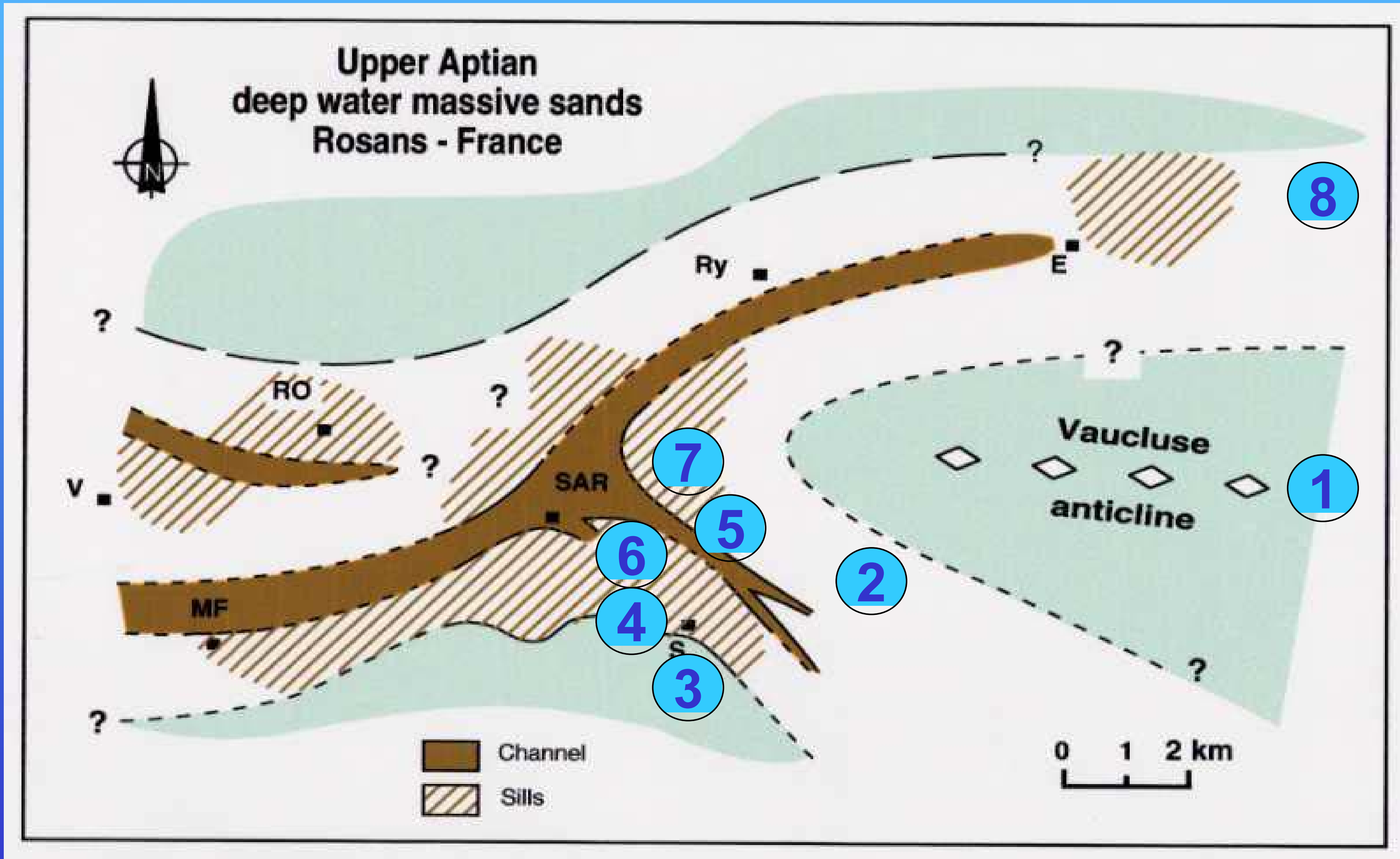


Friday 14th September 2001:

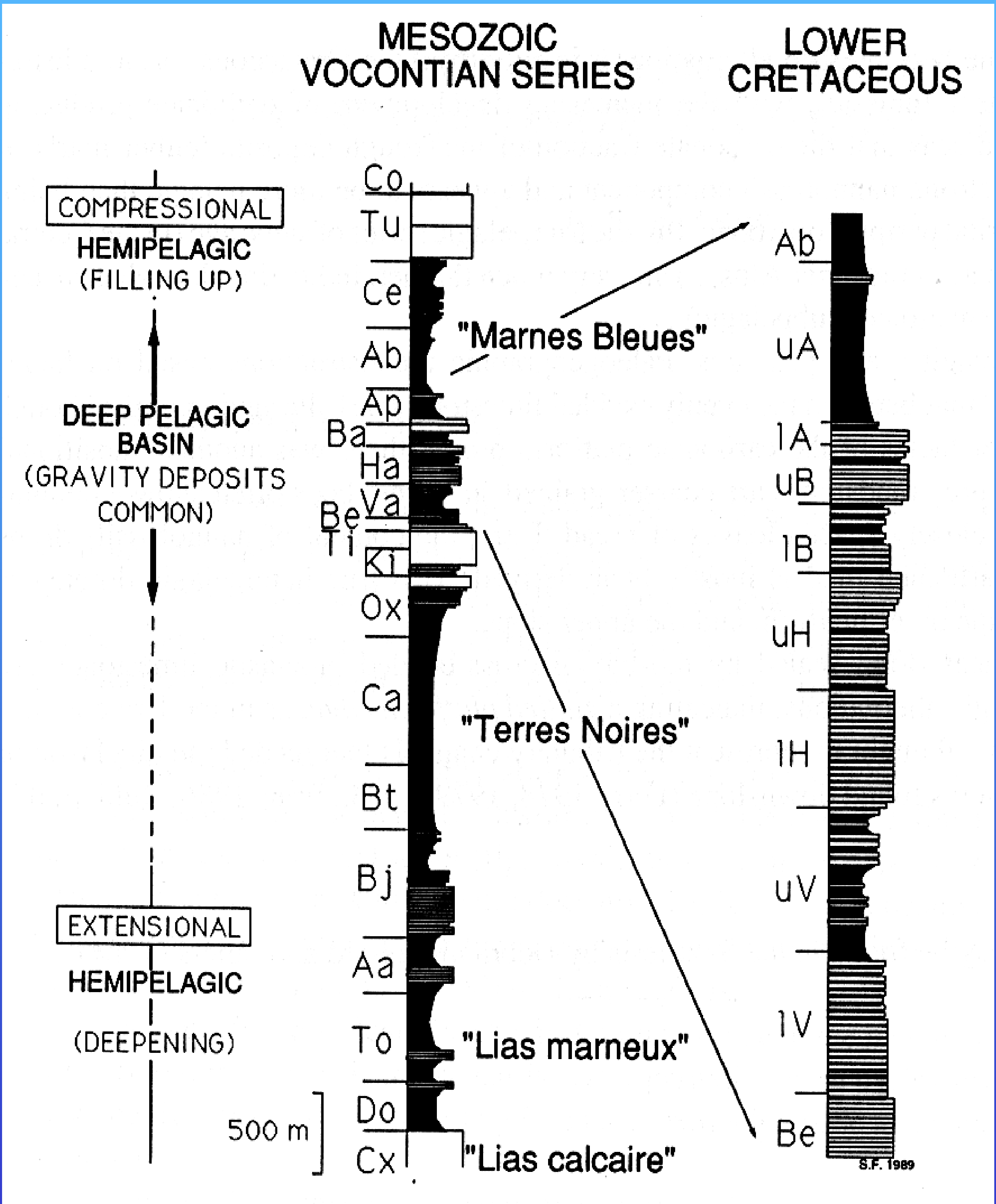
CLASTIC INJECTION EXAMPLES:

DEFINITION & CHARACTERISTICS

STATE OF THE ART.

A SILL-DOMINATED NETWORK.

Other topics: massive sands, erosive channels, slumps, sequence stratigraphy, topographic control on sand deposition, relationships between the injectites and their feeder



From S. Ferry (1984)

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Stop 1-1

Beaumont

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Rosans syncline - general view
Vocontian morphologies *versus* alpine structures

STOP 1: ROCHER DE BEAUMONT

THE VOCONTIAN DOMAIN

AN EXAMPLE OF TOPOGRAPHIC-CONTROLLED SLOPE SEDIMENTATION

La Barre Tithonique

During the next two days, the architecture of the will be dominated by the presence of a thick carbonate interval, present all over SE France: the "Barre Tithonique", a deep-water uppermost Jurassic and lower Cretaceous limestone. It is traditionally the first marker; an undergraduate student is able to recognise it when he/she starts mapping in the region, and could only be mistaken (by undergraduates) for the lower Cretaceous shelves around the Vocontian basin. We won't have this problem, since the trip will take us through the deep part of the basin only: all large limestone cliffs are Barre Tithonique, and what we are interested in sits above it.

General context

We shall be focusing on the earlier (Mesozoic) extensional phase (the Valaisan story of the Alps). It is actually quite comparable to the contemporaneous history of the North Atlantic, with a phase of rifting taking place from the late Palaeozoic up to the Upper Jurassic, and then mostly thermal subsidence of the margins.
The breakup took place in the Alps during the upper Jurassic, with production of oceanic crust more or less along the present axis of the Range (Monte Viso). The Nummulitic

Geodynamic setting of the Vocontian Basin

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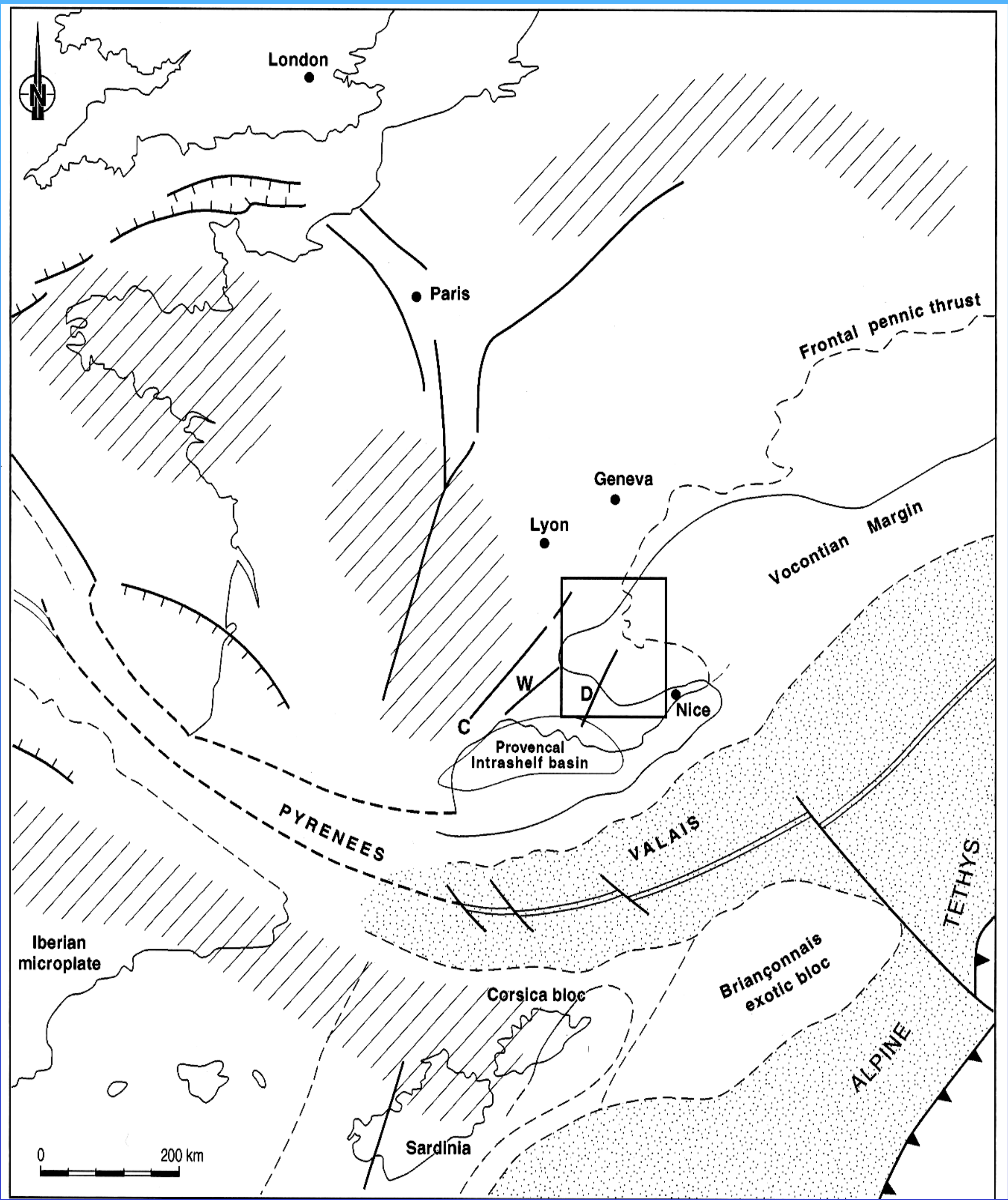

foredeep so called “Grès d’Annot” have been developed during the late compressional phase of the life of the Alps.

The Vocontian domain

During the Mesozoic, south-eastern France was thus the north-western continental margin of the opening Tethys ocean. The corresponding stratigraphy is shown on the Fig. 1.9 (DAY 1)
The Vocontian "basin" is defined from the lower Cretaceous (Uppermost Jurassic) to the Cenomanian as an area of deep-water sedimentation (pelagic or hemipelagic sediments) surrounded by platforms to the North (Vercors), to the West (Ardèche) and to the South (Provence). The eastern margin is not preserved, but the Vocontian basin was probably a gulf open towards the deep Tethys and its oceanic basin floor to the East. It is in that respect comparable to the Eocene North Sea which opened onto the North Atlantic. The main difference is that the tectonic activity was gentle, though continuous in the Vocontian basin. This is indicated by the long-lived submarine valleys which were the seat of active resedimentation throughout the lower Cretaceous.

The Lower Cretaceous sediments

The Cretaceous sediments of the Vocontian basin include numerous intervals of slumps, debrites and turbidites. These resedimented facies occur in discrete areas which define long lived submarine valleys.
The sediments are mostly carbonates and marls from upper Jurassic to lower Aptian; carbonate platforms at that time prograde from the margins of the basin and begin the infill. After that period comes the dominantly siliciciastic episode of the Apto-Albian with the déposition of the "Marnes Bleues" Formation (blue marls). This dominantly shaly series, several hundred metres thick, is interrupted locally by episodes of sand deposition. During that time, the sedimentation on the margins of this area is dominated by deposition of sandwaves related to shelf currents (Stop 1, Day 1). A map of the outcrops of the Aptian and Albian is shown on the field trip guide book cover.

Sedimentary setting

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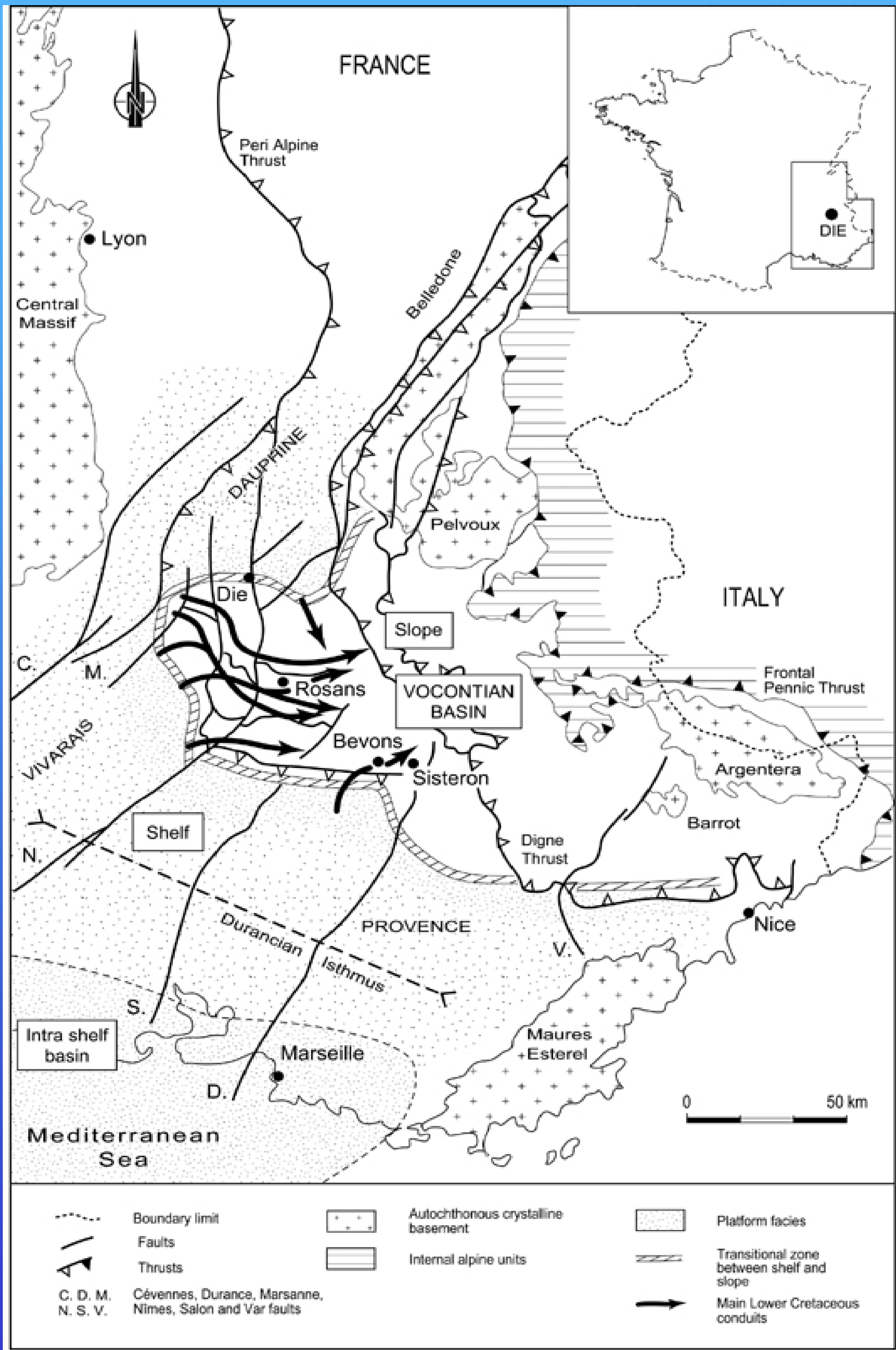

GENERAL STRUCTURE AND STRATIGRAPHY OF THE ROSANS AREA

PANORAMA ON THE SUBMARINE VALLEY-FILL

The Rosans area on which we will focus this day is located 15 km West of the Rocher de Beaumont. Rocher de Beaumont is on the crest of an anticline extending over some 15 km westwards: the Vaucluse anticline (between Stops 1 and 2). This gentle structure splits in two a major syncline, well marked by the “Barre Tithonique” which makes the sharp topographic crests to the North and South.

The panorama from the Rocher de Beaumont shows the general type of structuration of the Vocontian basin: anticlines and synclines.

At the same time, it provides a good overview of the local stratigraphy, from the Triassic which crops out in the heart of the Montrond diapir to the east up to the upper Cretaceous which makes the summit of Le Risou to the west.

To the East:

Triassic and Jurassic

Triassic evaporites (gypsum) crop out in the heart of the Montrond diapir, then Laragne and Upaix diapirs. They play a major role in defining the present structure, as well as the paleostructure.

The Lias does not crop out in the area

The Bajocian consists of marl-limestone alternations, up to 100 m thick

The Terres Noires Formation encompasses the Bathonian, Callovian and Oxfordian. They consist of thick (about 2000 m), monotonous dark shales (with little source rock potentiel, though, the black colour is due to the presence of pyrite). Their thickness can be increased by tectonics.

The Upper Jurassic is marked by a progressive increase of carbonate content, which leads up to the "Barre Tithonique", a thick (100 to 150 m) massive carbonate formation which underlines tectonic structures in the region.

To the West:
Cretaceous

The plunging axis of the Vaucluse anticline to the west brings to the outcrop the whole Cretaceous series, from the Berriasian up to the Turonian which makes the Risou summit 15 km to the west.

Above the "Barre Tithonique", the series gets progressively marly with the Valanginian marls (300 m), then records an increase in carbonate content with the development of a second massive limestone during the Barremian and lower Aptian (barre barrêmo-bédoulienne, 50 m). This limestone episode is coeval to the main development of the Vercors carbonate platform to the north. The rest of the Aptian and Albian are characterised by a dominantly shaly sedimentation (Marnes Bleues Formation, 800 m) interrupted by siliciclastic episodes. The uppermost part of the Cretaceous visible in the syncline is the 200 m thick Cenomanian-Turonian limestone of the Risou.

We will focus on the western part of the syncline, more precisely where the Vaucluse anticline terminates, leaving one simple syncline. In that area, the core of the syncline is filled by a thick, massive sandbody, Uppermost Aptian in age, deposited in deep water about 40 km from the shelf (La Lance area, cf back cover of your guide book)

Although most of the structure is recent, the fact that Aptian sand deposition is restricted to the synclinal axes indicates that it was already initiated at the time of deposition.

Thanks to the tectonic style of the area, they are widely scattered over the basin, which allows precise paleogeographic reconstructions on which to base the interpretations.

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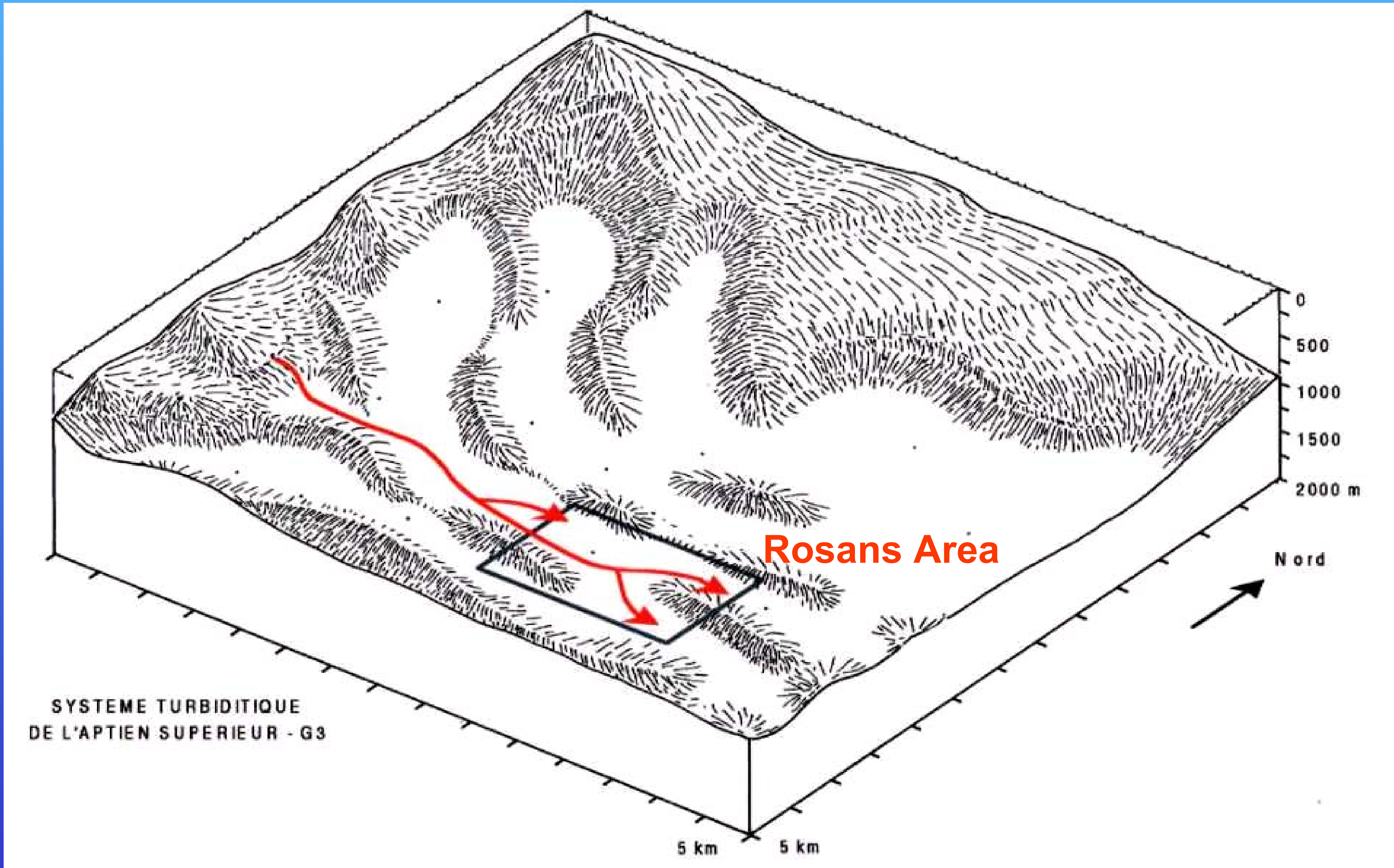
Upper Aptian Turbiditic « Massive » System



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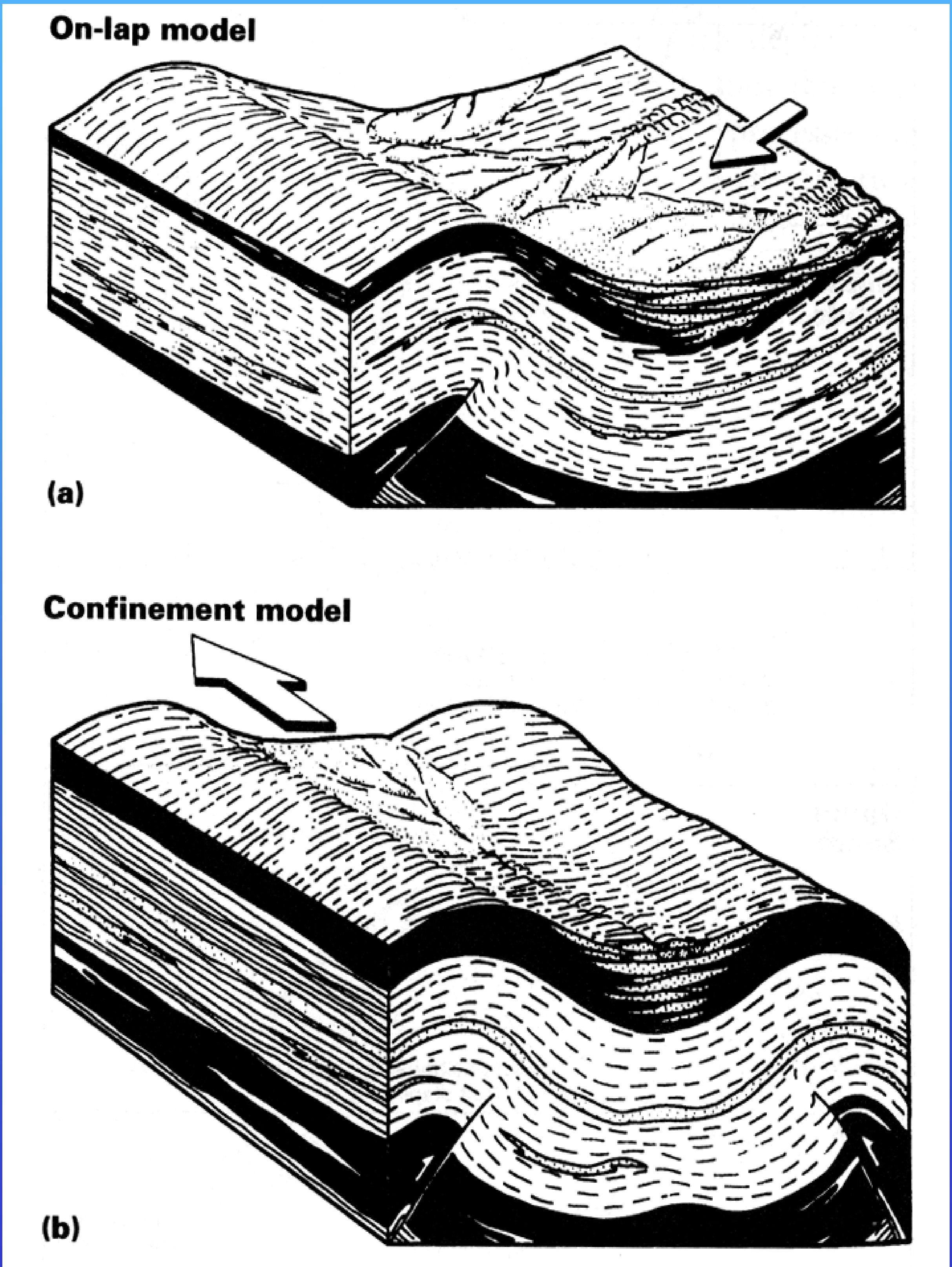
Scott & Tillman (1981) 's model



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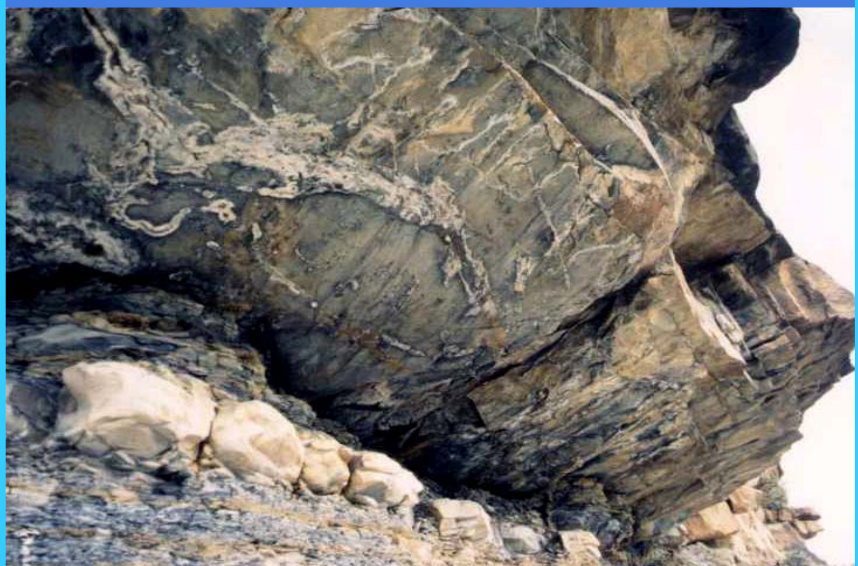
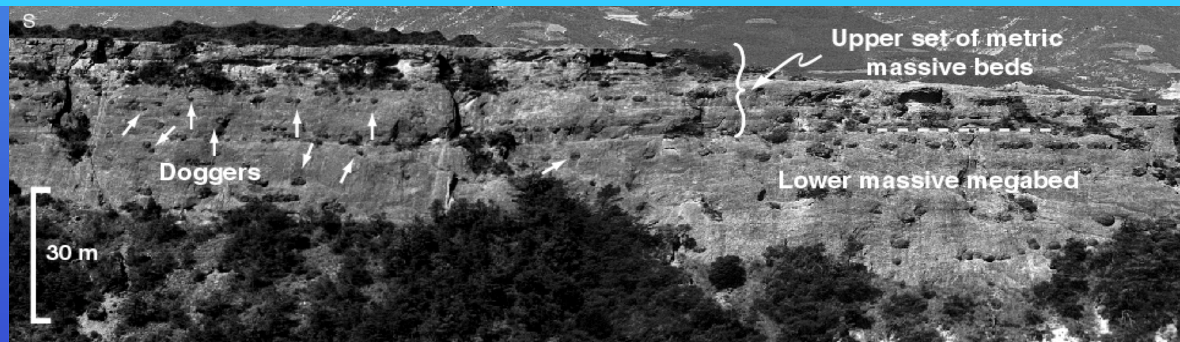
The Aptian « Blue Marls » Formation





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Stop 1-2

Sorbier

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Litho-stratigraphical setting

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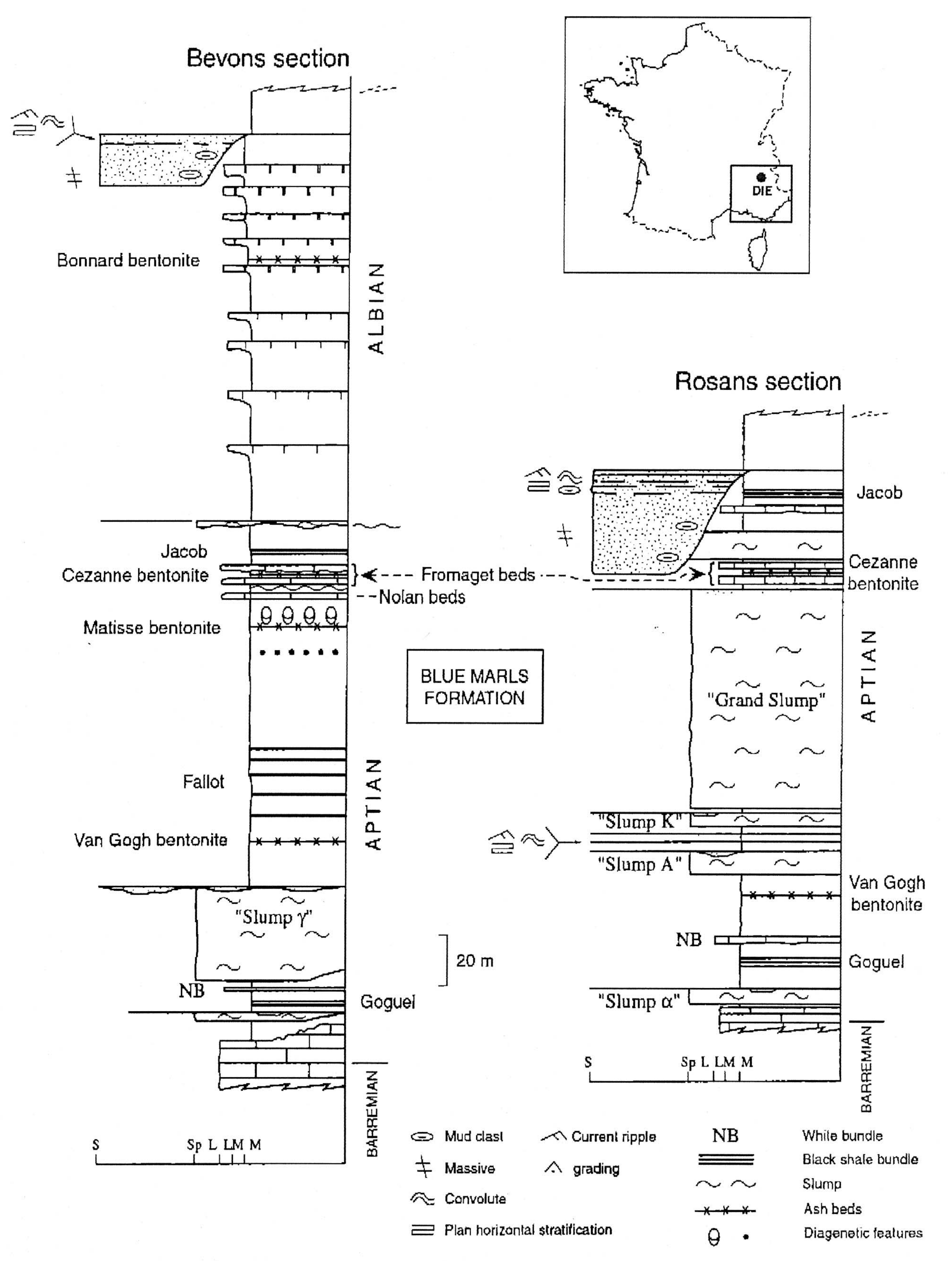


STOP 2: SORBIERS

THE APTIAN “MARNES BLEUES” FORMATION AND THE MASSIVE SANDS OF ROSANS

Two points here: a closer view on the stratigraphy of the “Marnes Bleue”, and a first feeling of what massive "turbidite" sands can be. The rest of the day will be spent within a few km of this stop.

THE APTIAN “BLUE MARLS” FORMATION

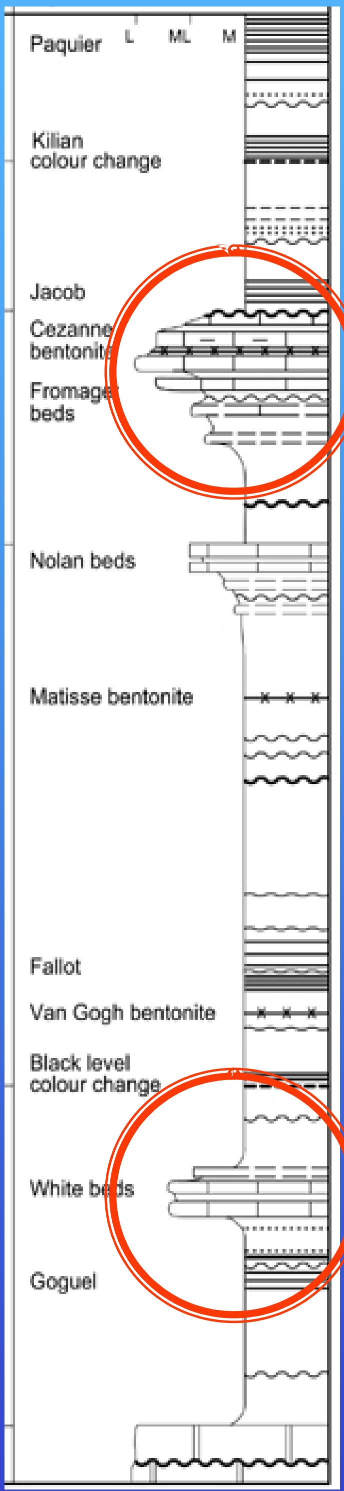
About 150 m of "Marnes Bleues" can be seen from this point, above the Barremo-Bedoulian limestone. They are capped by the massive Rosans sandstone complex which will be studied in detail during the day.

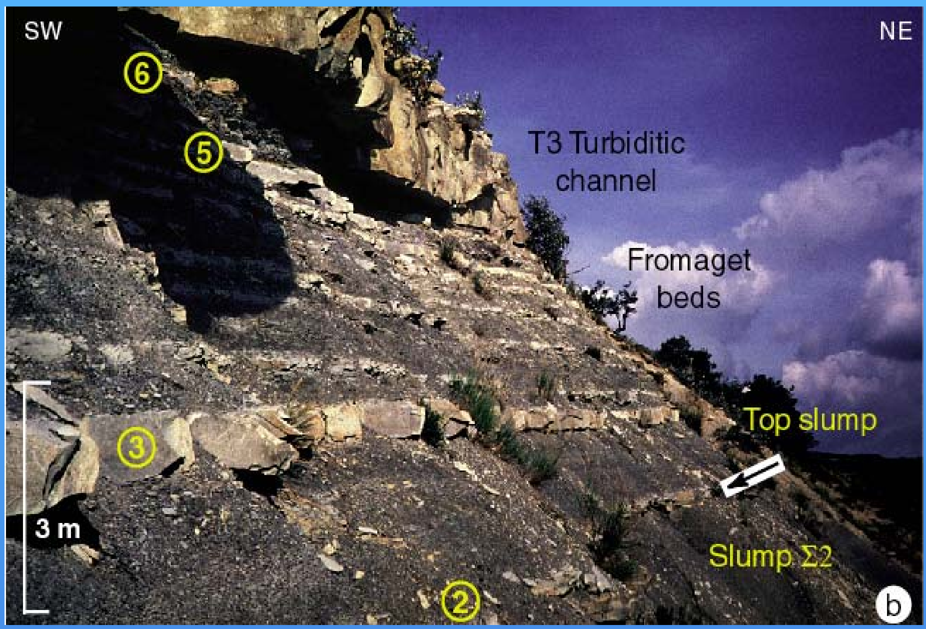
Before focusing on the Rosans sandbody, it is worth having a look at the other resedimented facies observable from Sorbiers:

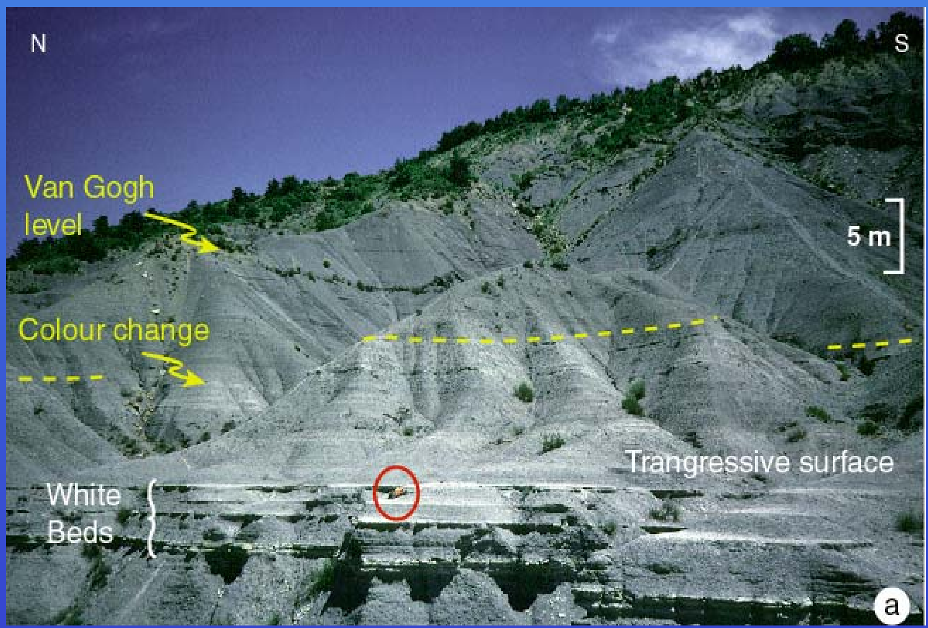
* The White beds, a key-bundle in the Vocontian area

* An older sandy system T1 (Middle Aptian) develops in the southern part of the outcrop, and is made up of standard turbidites. We discuss the sequential organization of the Aptian succession.

Some key-beds







Middle Aptian depositionnal sequences

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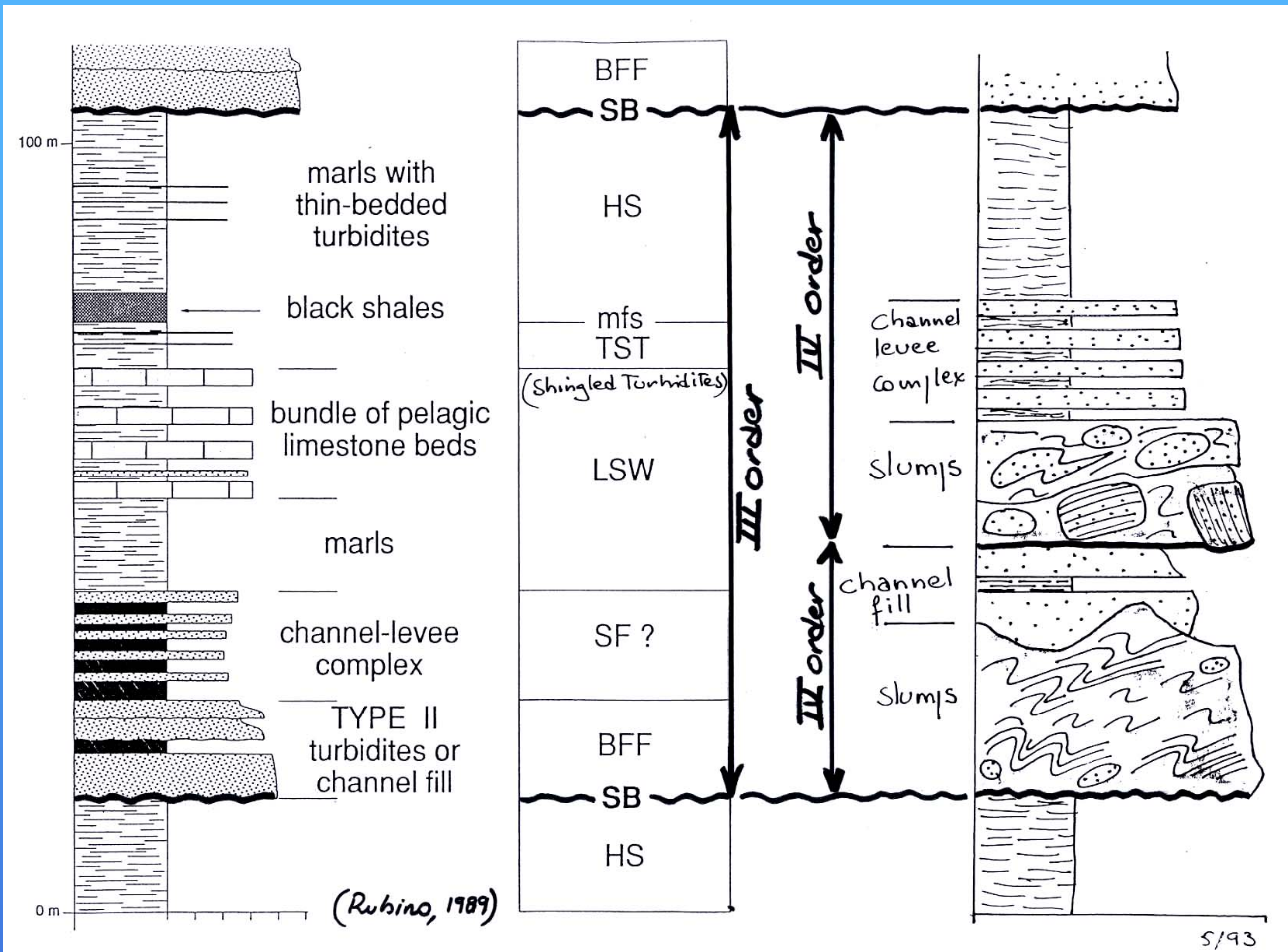

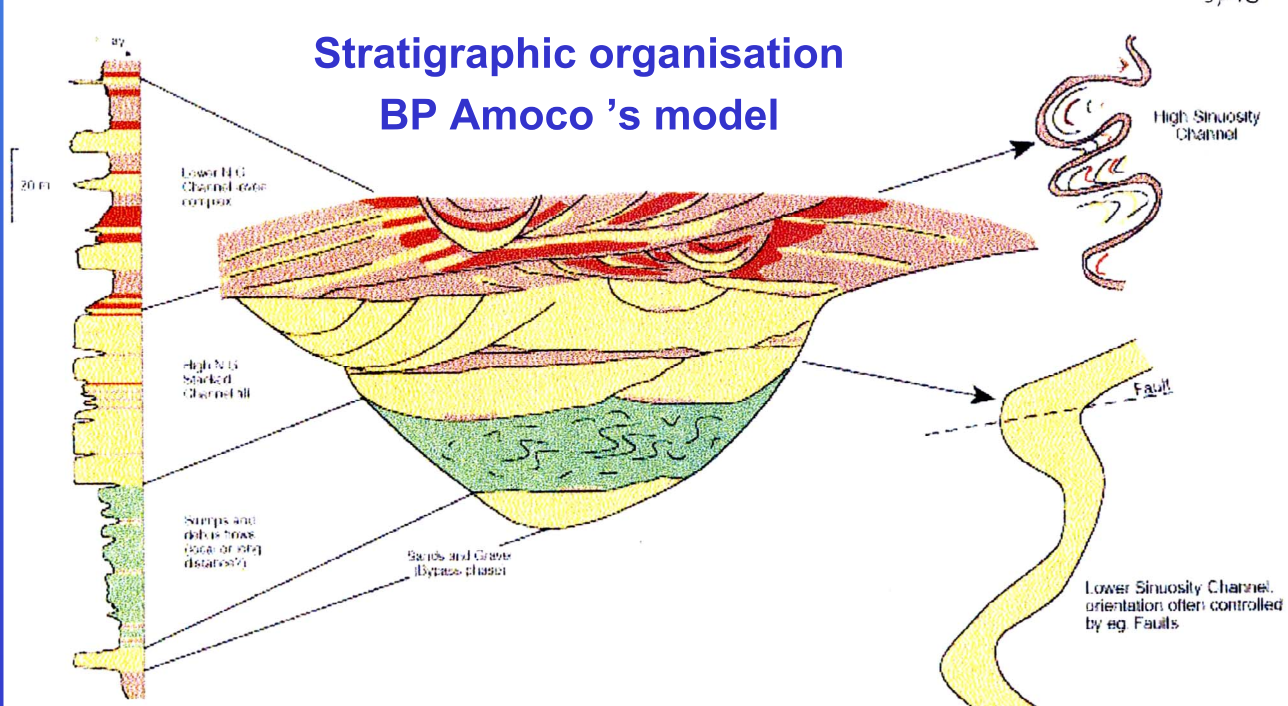

Stratigraphic organisation
BP Amoco 's model

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General view
of Upper Aptian Rosans sandbodies

* Between this lower sandy system and the upper massive one, most of the "Marnes Bleues" are actually made of one large-scale slump which we will see several times during the day. This "Grand Slump" is up to 80 m thick, locally disturbed enough to be recognised as such, but often quiet enough to be mistaken for *in-situ* sedimentation. This should be kept in mind when looking at cores or logs: it is very easy to miss a glided mass of sediment in a well, especially when it is a big thing. We shall see later during the day a good outcrop of “Grand Slump”.

THE ROSANS SAND BODY

The stop also gives a good general view of the Rosans sandbody. It appears really massive, without any shale interbed. The maximum thickness is 40 metres at Serre d’Autruy, at les Aigrets hill (on your left) this thickness decreases to 5 metres. The sandbody can be followed for about 5 km N-S. The lenticular shape of the Rosans sandbody results from a combination of recent erosion and differential compaction between the sandstones and shales, without it being possible to decide whether it was gently mounded or boringly flat at the time of deposition.

In more detail, it is possible to see locally (binoculars recommended) the erosion on the margin of the sandbody, which stresses its channelized character. Three main erosive phases and three infilling phases can be distinguished. Each infill began by a megaturbidite body (10 to 30 m thick). These large oblic erosive surfaces looking like point bar features: we dicuss this organization
(Stop 6, this day)

Mapping of Upper Aptian sandstones

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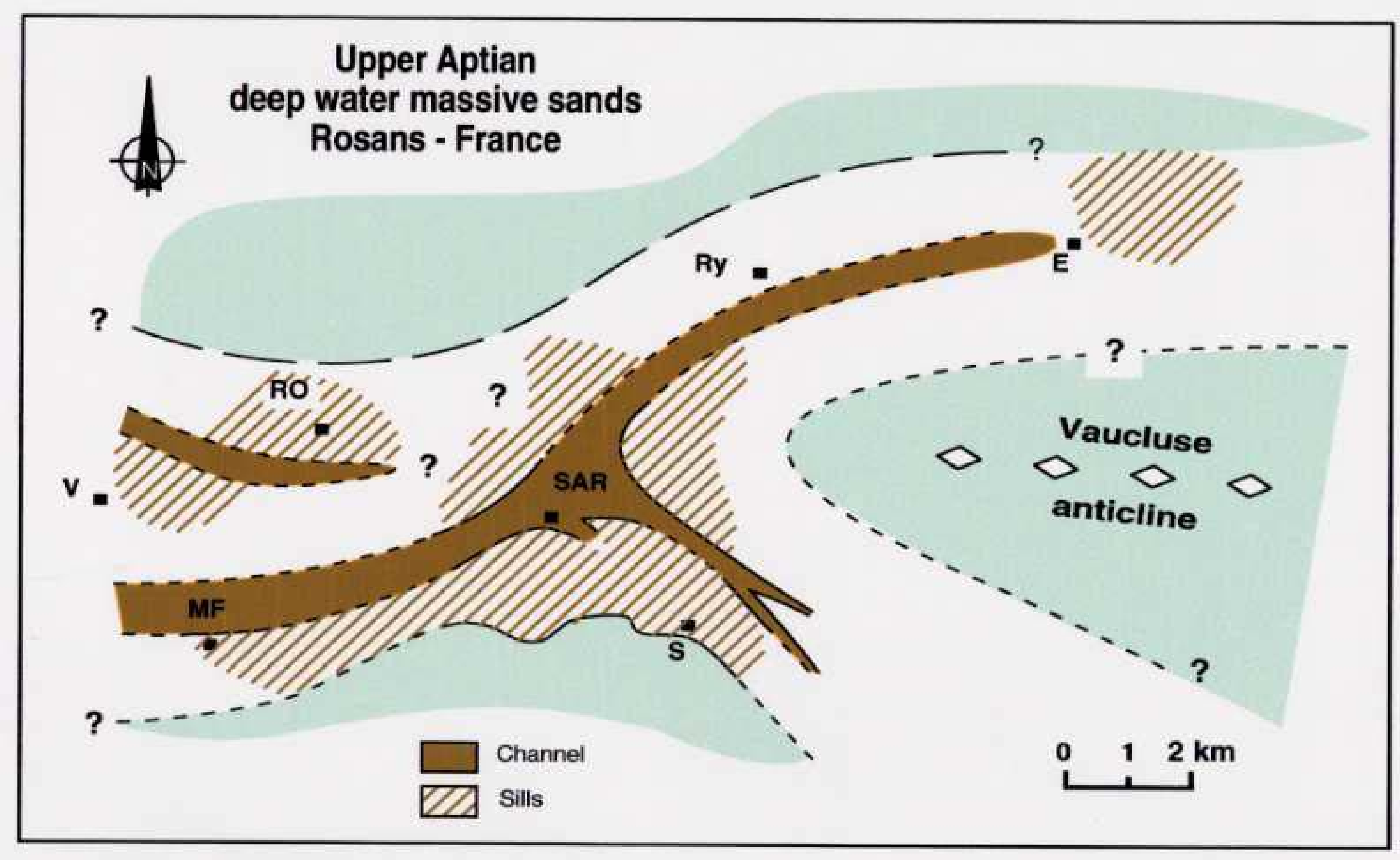

The « Vocontian massive sand » model

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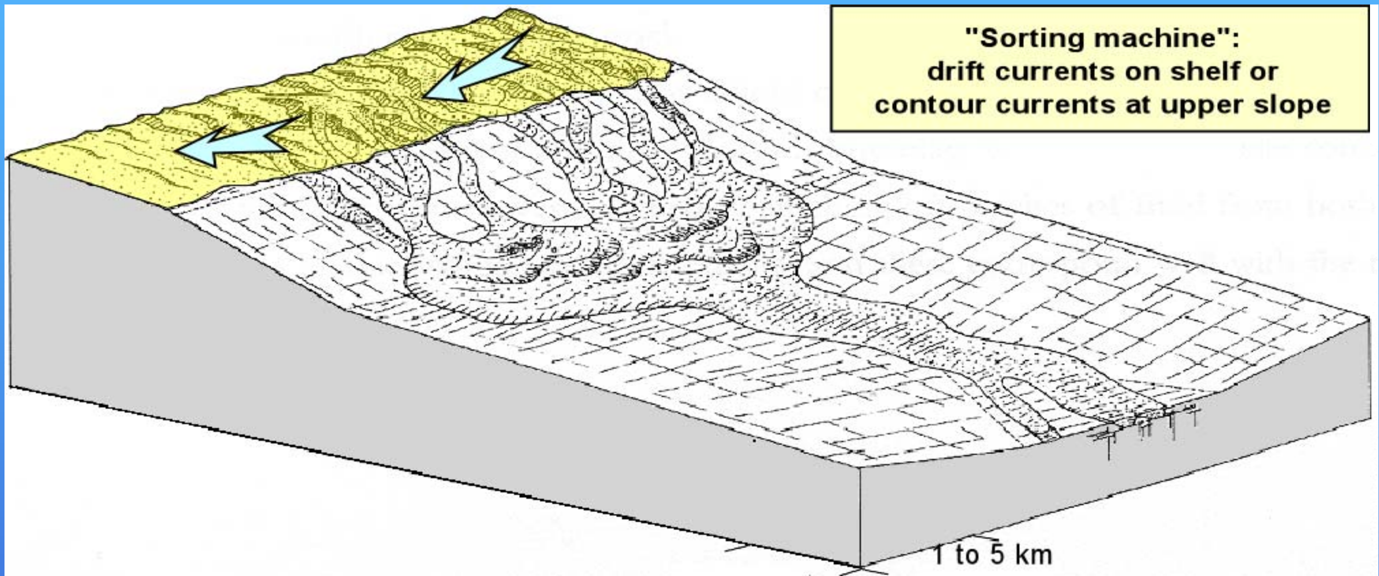

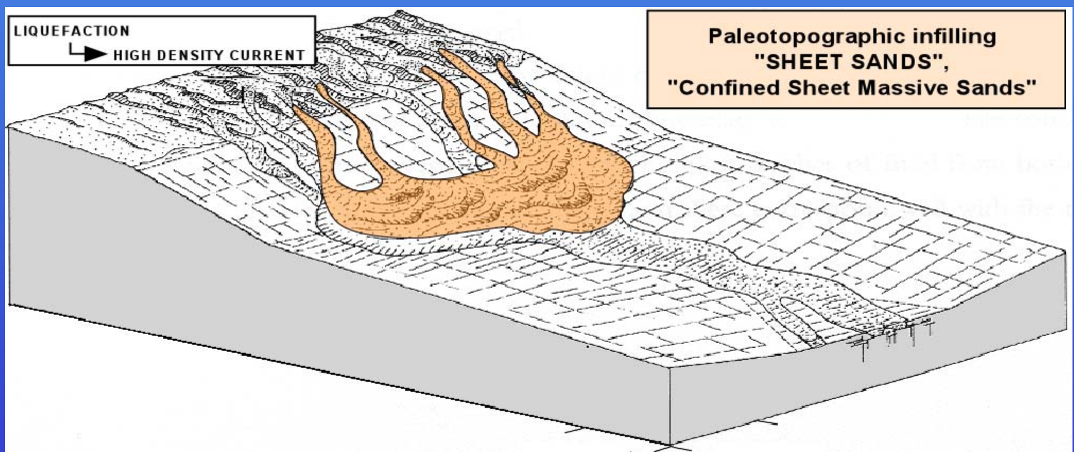

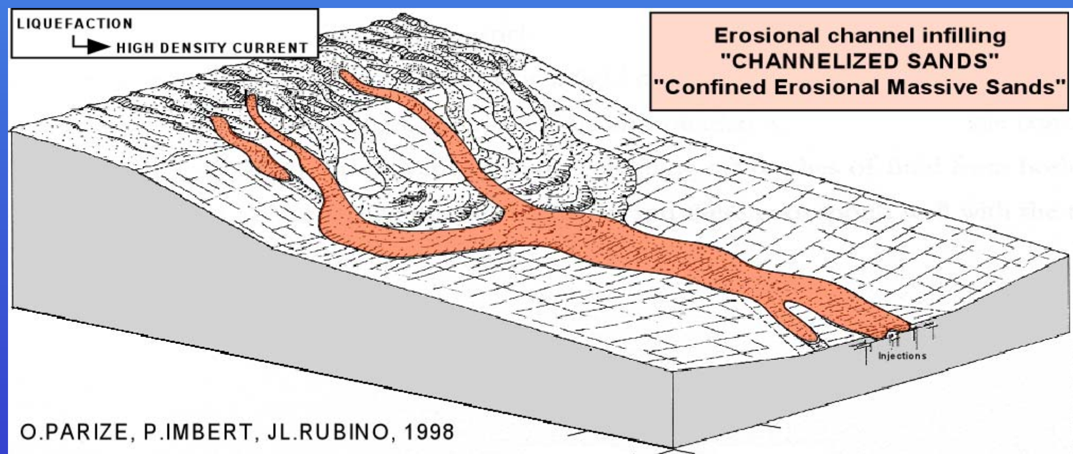

Upper Aptian key-beds and sedimentary bodies

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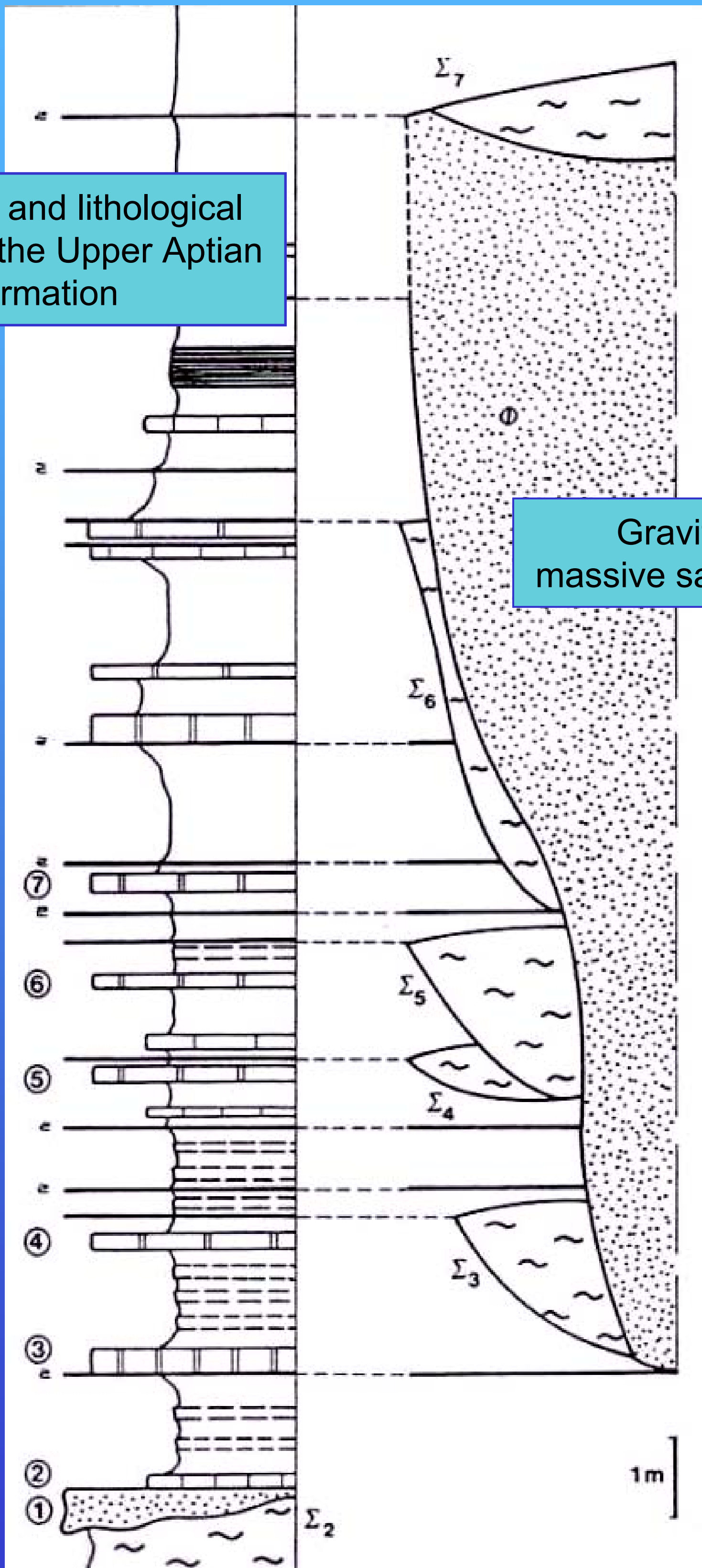

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Stop 1-3

Sironne

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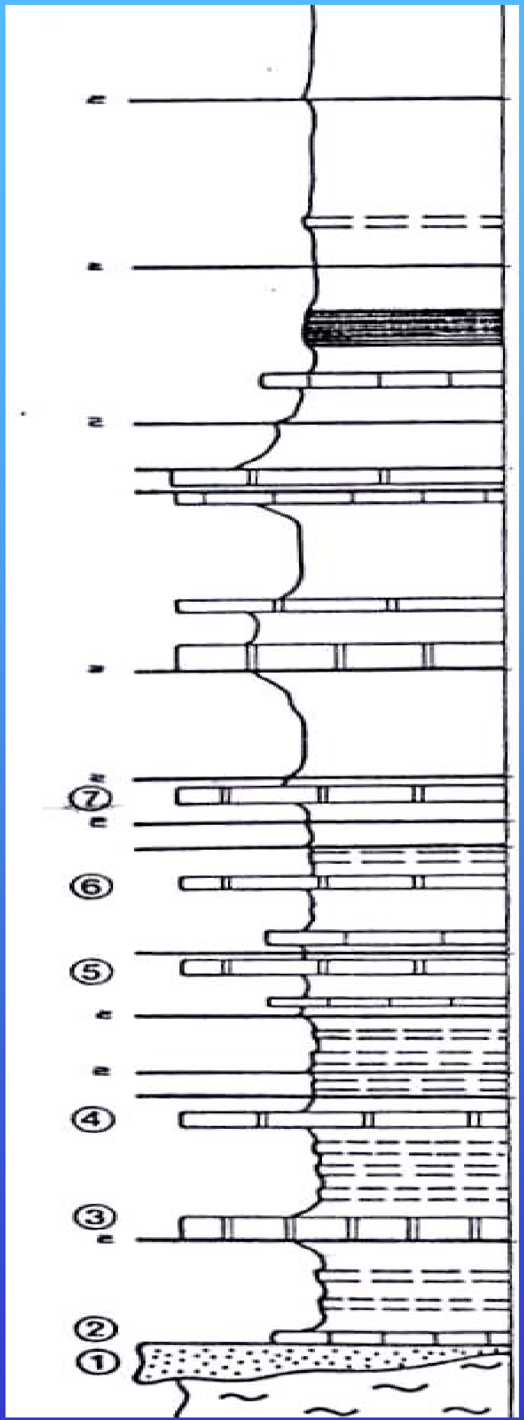
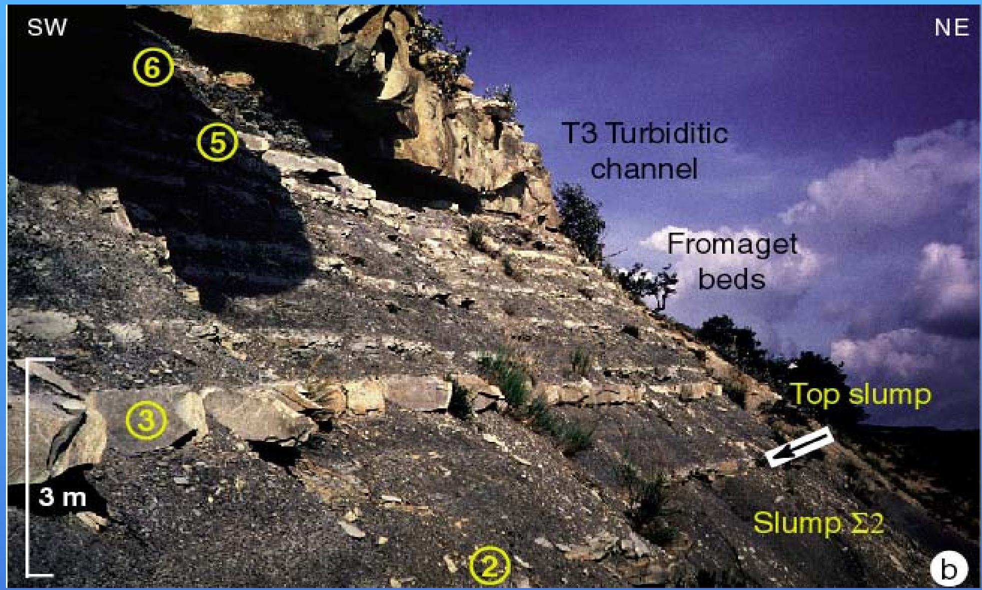
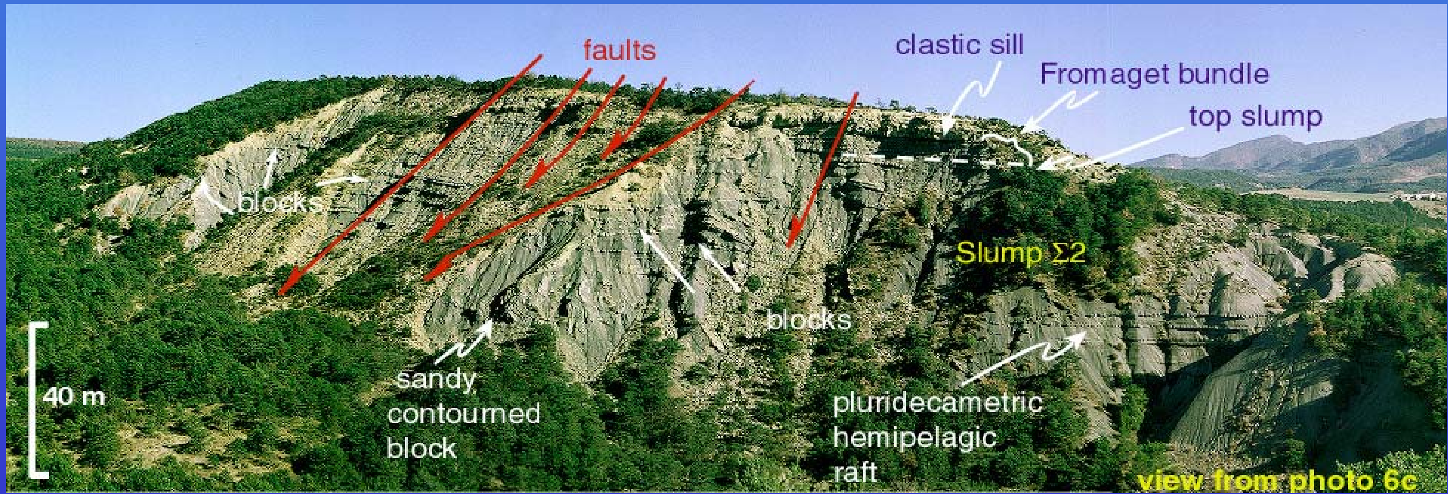
Fromaget bundle and its key beds







STOP 3: SIRONNE

THE UPPER APTIAN FORMATION

SOME KEY BEDS

A guide: the key beds of the Fromaget bundle

Several limy and marly-limy beds can be correlated on the western par of the vocontian domain: the main ones are key beds n°3, n°5, n°6 and n°7.

And here, we don't observe sandstone bed in the Fromaget bundle

The "Grand Slump"

The "Grand Slump" (Friès, 1987) is the most important slump in the Aptian-Albian formation. Its thickness is up to 100m.

A first point of view on the Toulay cliff from Sironne

Toulaye cliff is the southern end of la Serre d'Autruy. It shows the southern (right) bank of the channelized massive sands injected by a complex network of sills and dykes. This stop allows a good view of the southern part of the Toulaye cliff.

Toulaye Cliff - General view (1)

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Stop 1-4

Faysses

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Your drawing of Toulaye cliff (1)

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STOP 4: LES FAYSSES

GENERAL VIEW OF TOULAYE CLIFF (SOUTHERN PART):

Toulaye cliff is the southern en of the serre d'Autruy. It shows the southern (rigt) bank of the channelled massive sands injected by a complex network of sills and dykes. This stop allows a good view of the southern part of the Toulaye cliff.

In order to better understand the complexity of this outcrop, a (quick) drawing is required. It will help along with the analysis and progressive characterization of the different geometry of the sandy body this afternoon.

Binoculars (Swiss hammer) may be helpful.

Toulaye Cliff General view (2)

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Stop 1-5

Aigrets

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Toulaye Cliff - General View (3)

Next stop
after the lunch

STOP 5: LES AIGRETS WEST

GENERAL VIEW OF
TOULAYE CLIFF (NORTHERN PART):
THE SILL DOMINATED NETWORK AND ITS
RELATION WITH THE FEEDER

The “Point-of-view” on the Toulaye cliff and its sandstones bodies

This stop allows a good view of the northern part of the Toulaye cliff. You can continue your drawing of the Rosans sandbody. Binoculars (Swiss hammer) may be always helpful. You can find the channel and the relation berween the injected bodies and their feeder.

A guide: the key beds of the Fromaget bundle (again)

The “Grand Slump” (again)

To climb to the top of the hill we cross te Grand slump and the key beds of the Fromaget bundle. The Aigret cliff provide very good exposures of the Grand slump : this outcrop was previously described by Goguel en 1938.

Provençal Pic Nic
with the support of Genevière and Didier from
“Ferme Auberge Vaujeala”

Your drawing of Toulaye cliff (2)

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Your drawing of Toulaye cliff (3)

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Lunch

Along Aigues river

***Provençal* Pic Nic**
with the support of Genevière and Didier from
"Ferme Auberge Vaujeala"

Diagnostic features for sills

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Stop 1-6

Toulaye cliff

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An example of sill-dominated network

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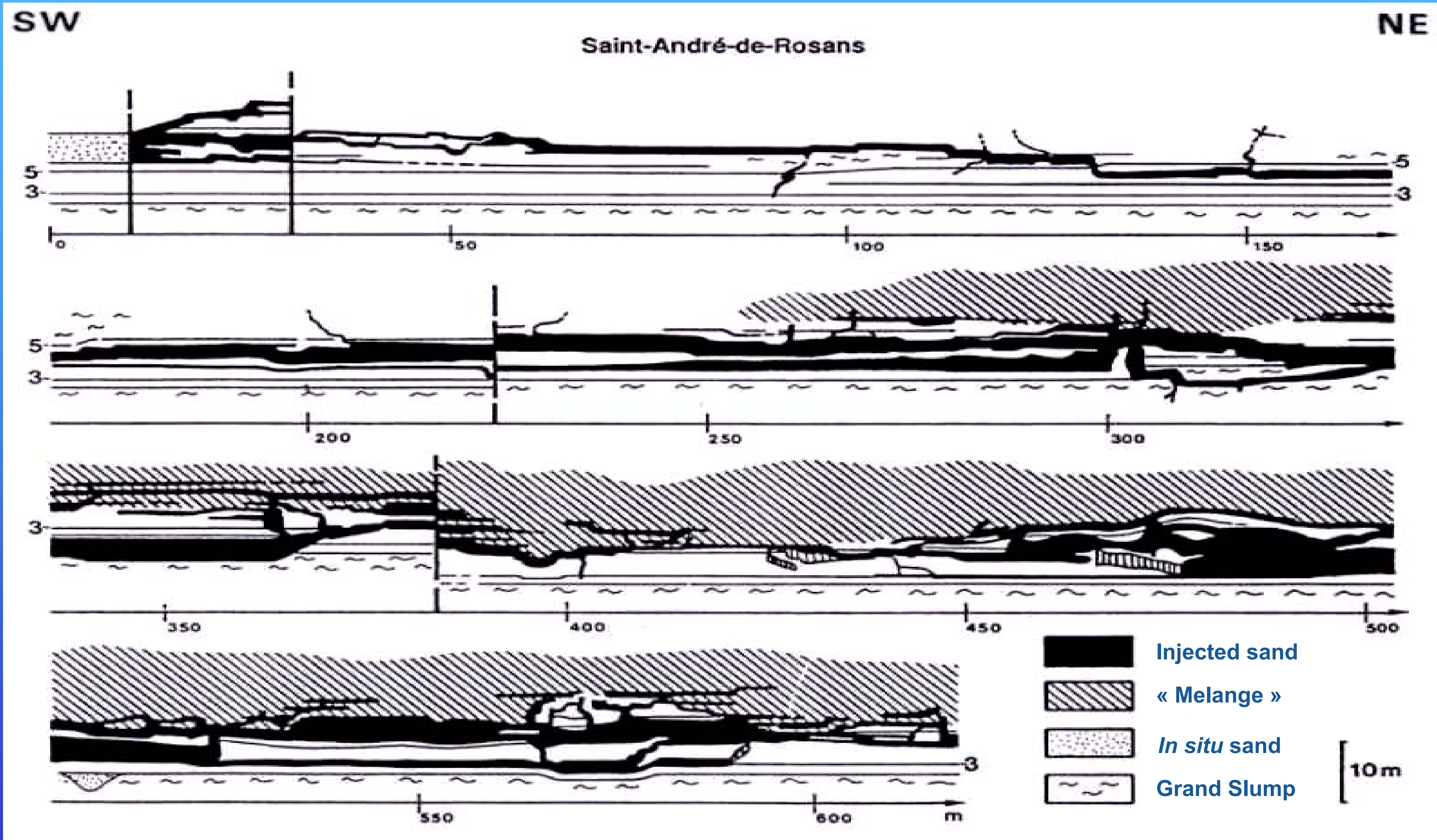


STOP 6: TOULAYE

WALKING ON
THE CLASTIC SILLS AND DYKES

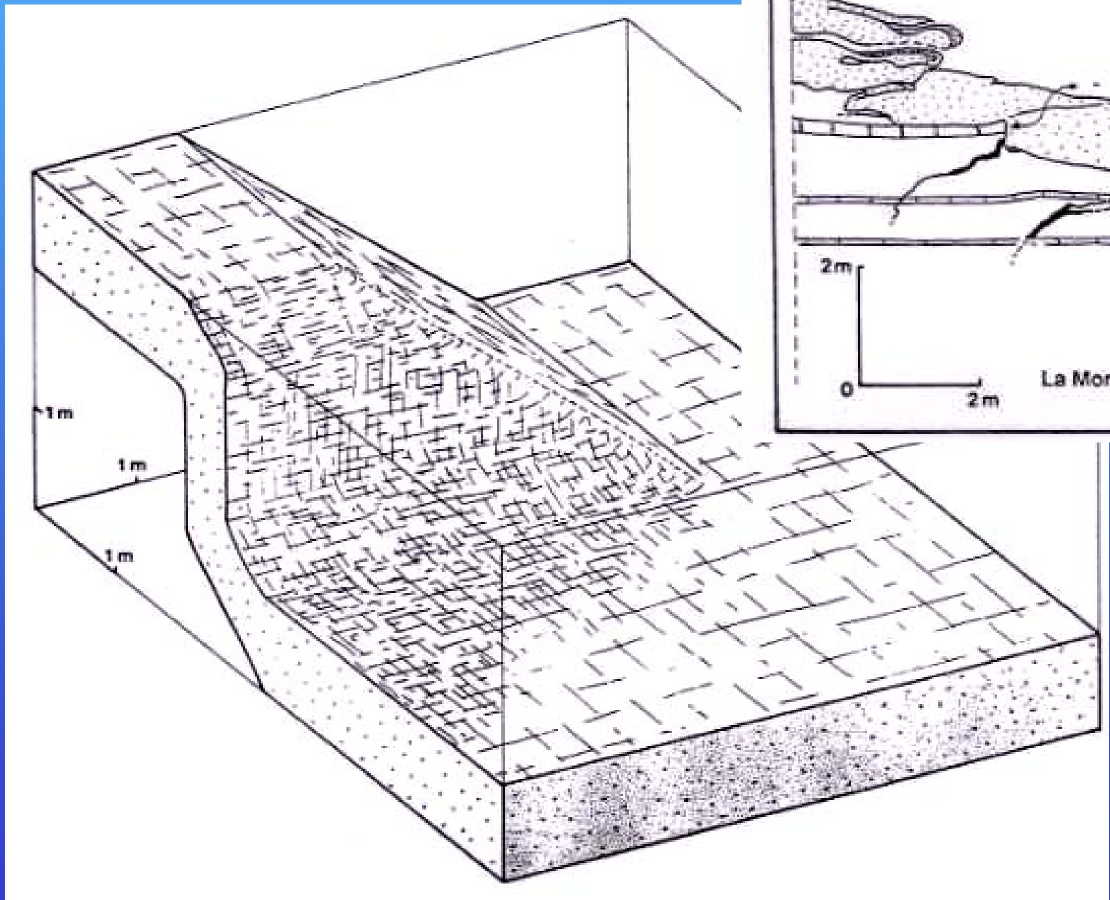
Geometrical characteristics of the sills

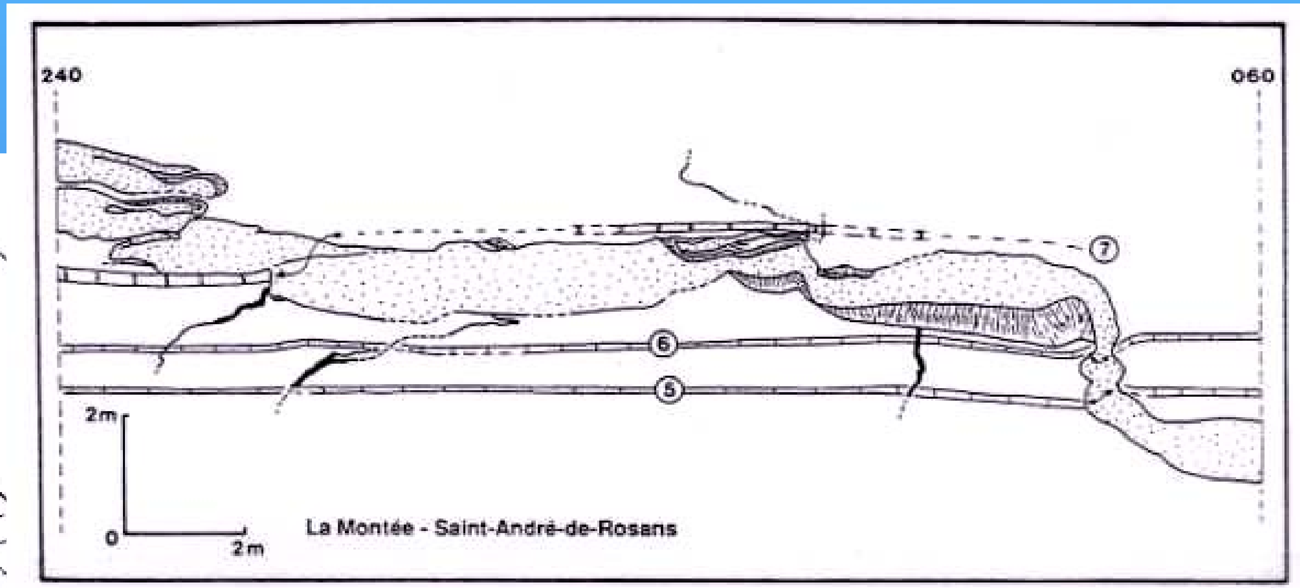
The sills could be mistaken for massive sand beds, their main distinctive character appears when they "jump" from one level to another. They can show frondescent casts at the contact with the shales. They do not show tractive sedimentary structures, but water escape features are locally present (dish structures).

The thickness of the sills and dykes varies from a few centimetres to several metres. The maximum thicknesses are observed close to the feeder channel, and the cumulative thickness of the injections decreases away from there. In Saint-André-de-Rosans, sills can be traced up to 2.5 km from the margin of the channel.

The sills are injected from the bottom part of the sandbody, about 30 m below the sea floor at the time of the injection. Away from their feeder channel, they are injected higher and higher up in the series through successive jumps via dykes.

Clastic sill geometric organisation





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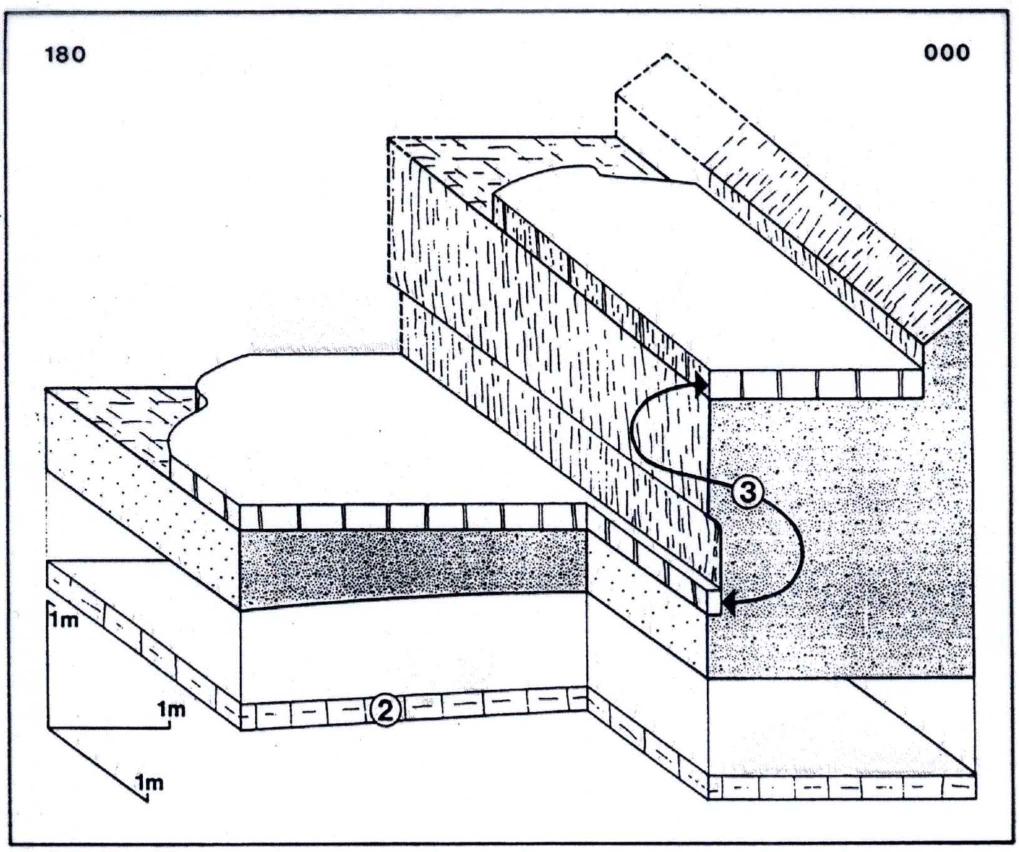
Clastic sill geometric organisation



Sills and dykes relations

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Imperial Wall:
a dyke parallel to the outcrop trend

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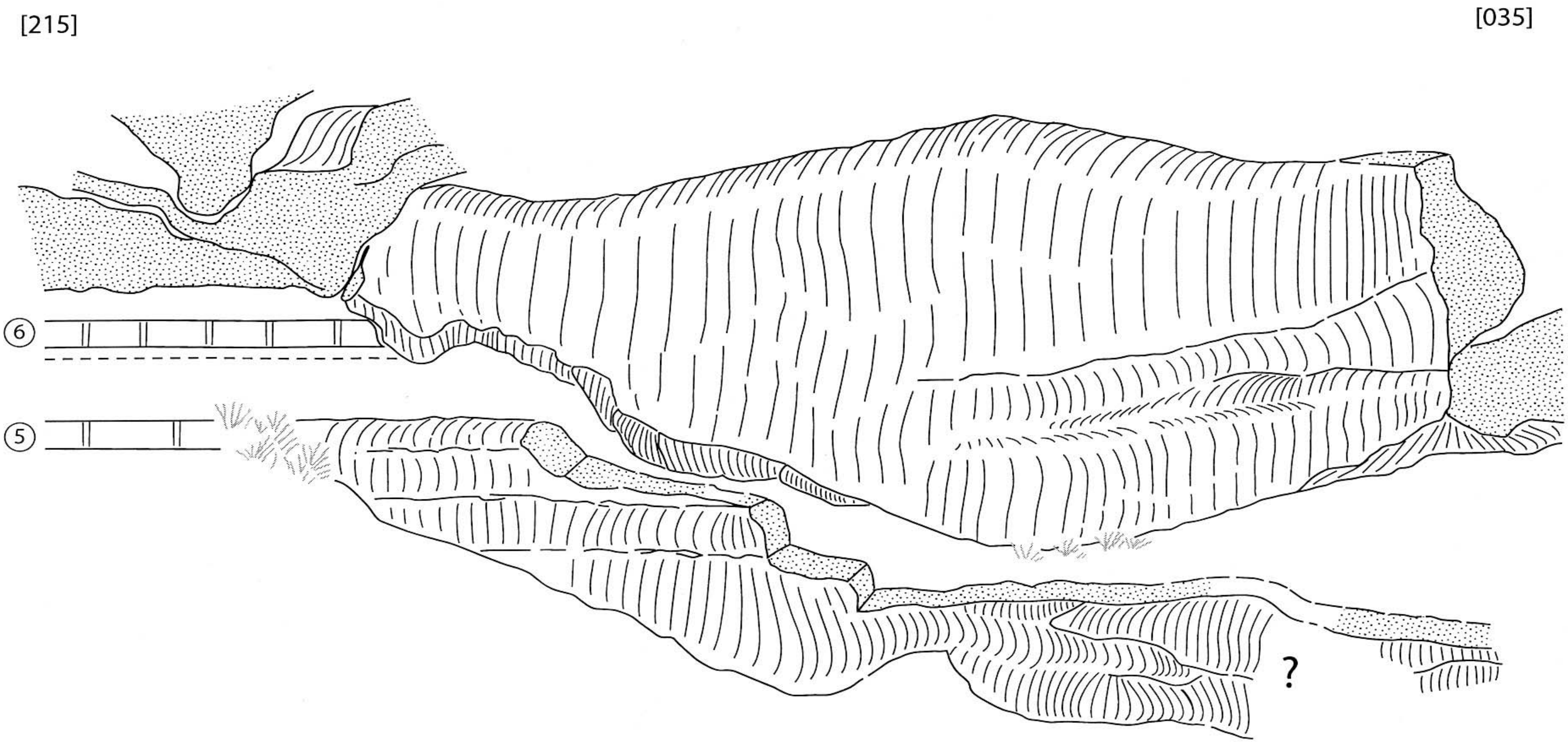

Clastic injections network

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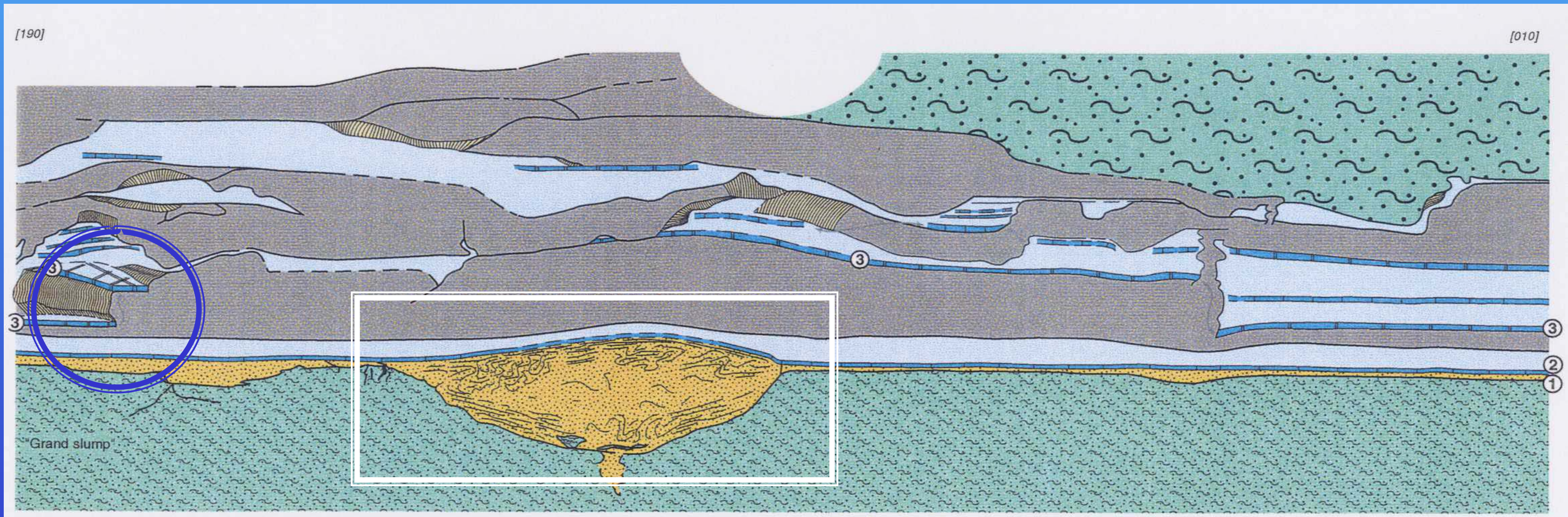
PS : Beware of the lenticular features as « Imperial wall »



Stratigraphic location of injected levels

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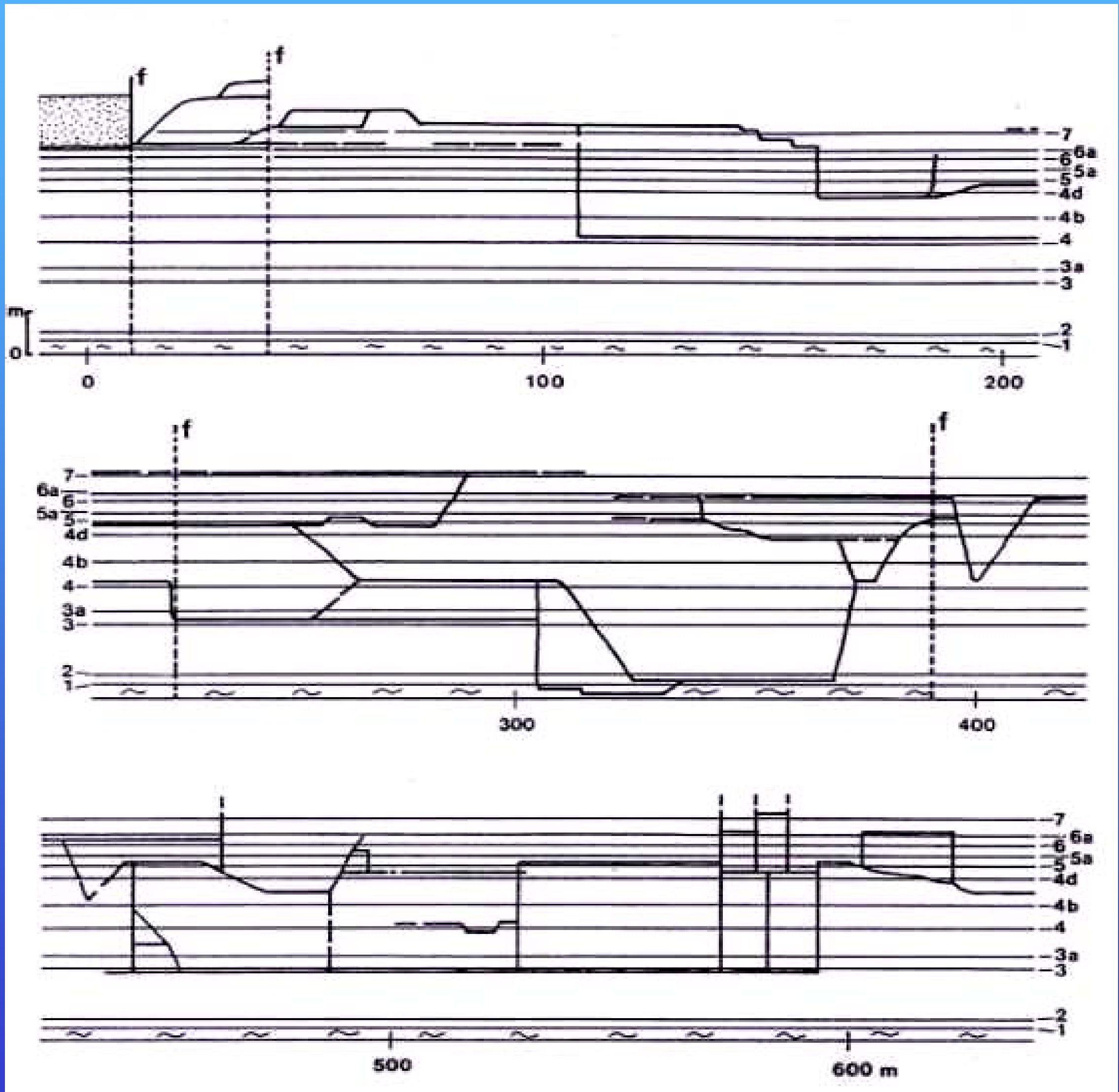

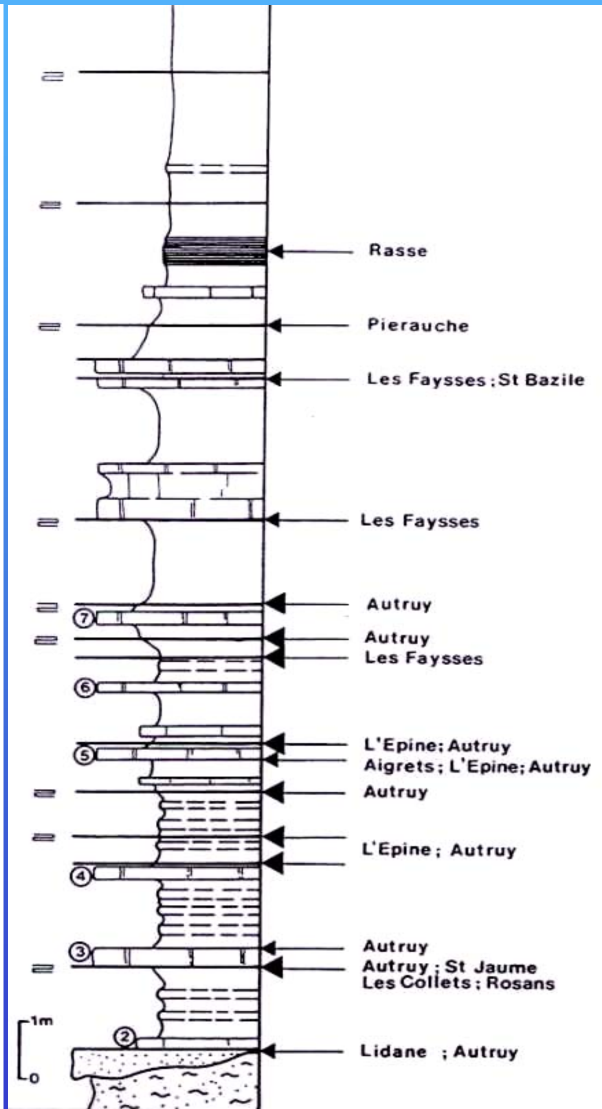

The dimensions of the injectite network

A detailed volumetric approach conducted in the Saint-André-de-Rosans sector, up to 10 square kilometres, indicated that the injected sand volume (southern part of the network) corresponds to 8.10^6 m^3 and the channel volume to 35.10^6 m^3, so slightly less than 20% of the sand volume is injected and if the northern network volume is volume-similar, this ratio would be up to 30%. These dimensions are similar to those described in subsurface examples, e.g. Balder field (Jenssen *et al*., 1993) or Alba field (Newton & Flanagan, 1993, McLeod *et al.*, 1999).

The injectite facies, relationship with the litology of the host formation

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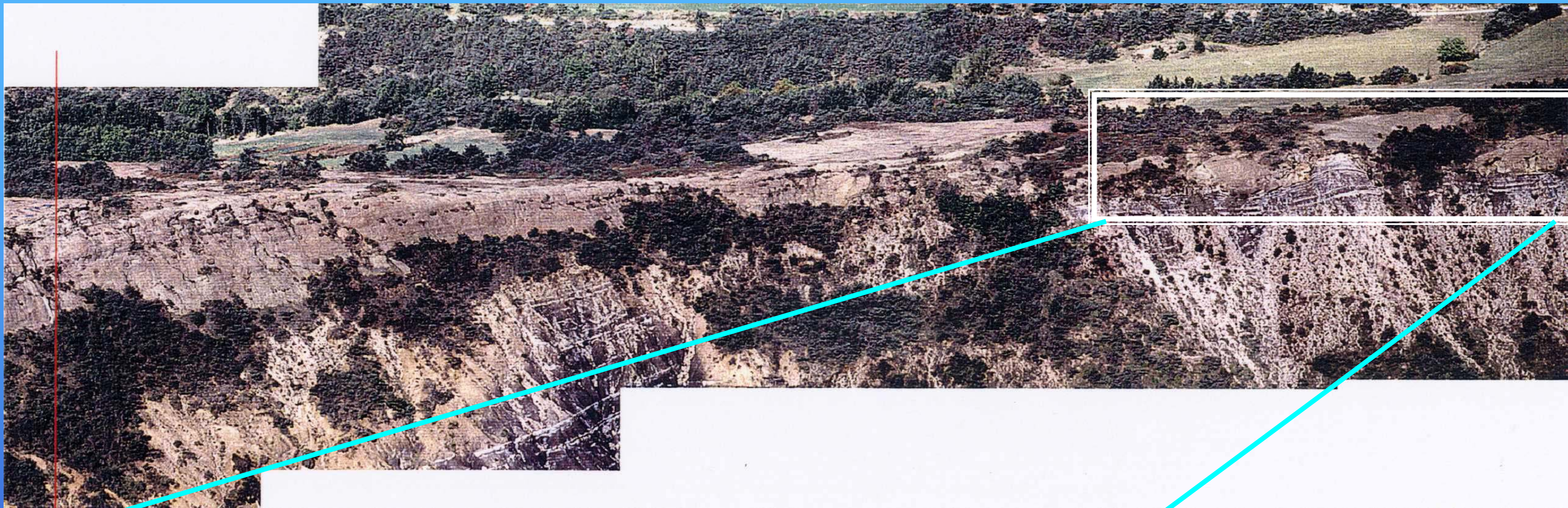
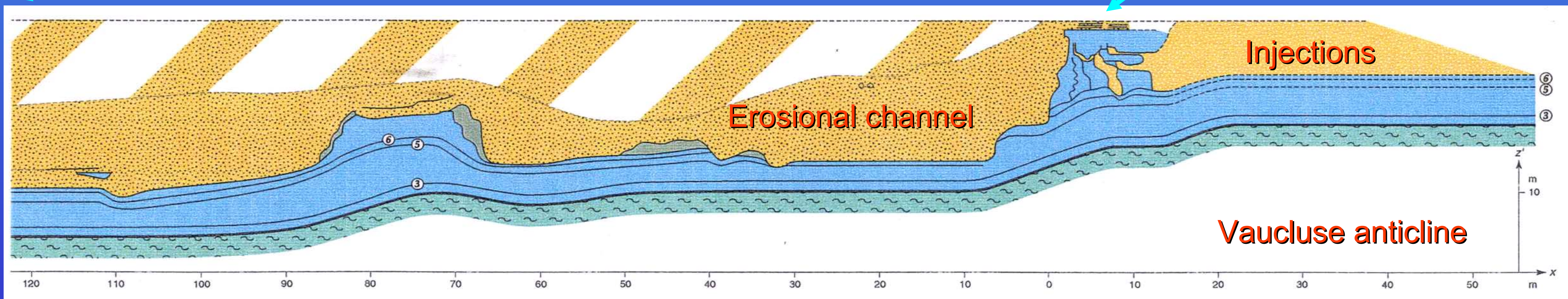
Injections developped in the channel margin





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Stop 1-7

Serre d 'Autruy

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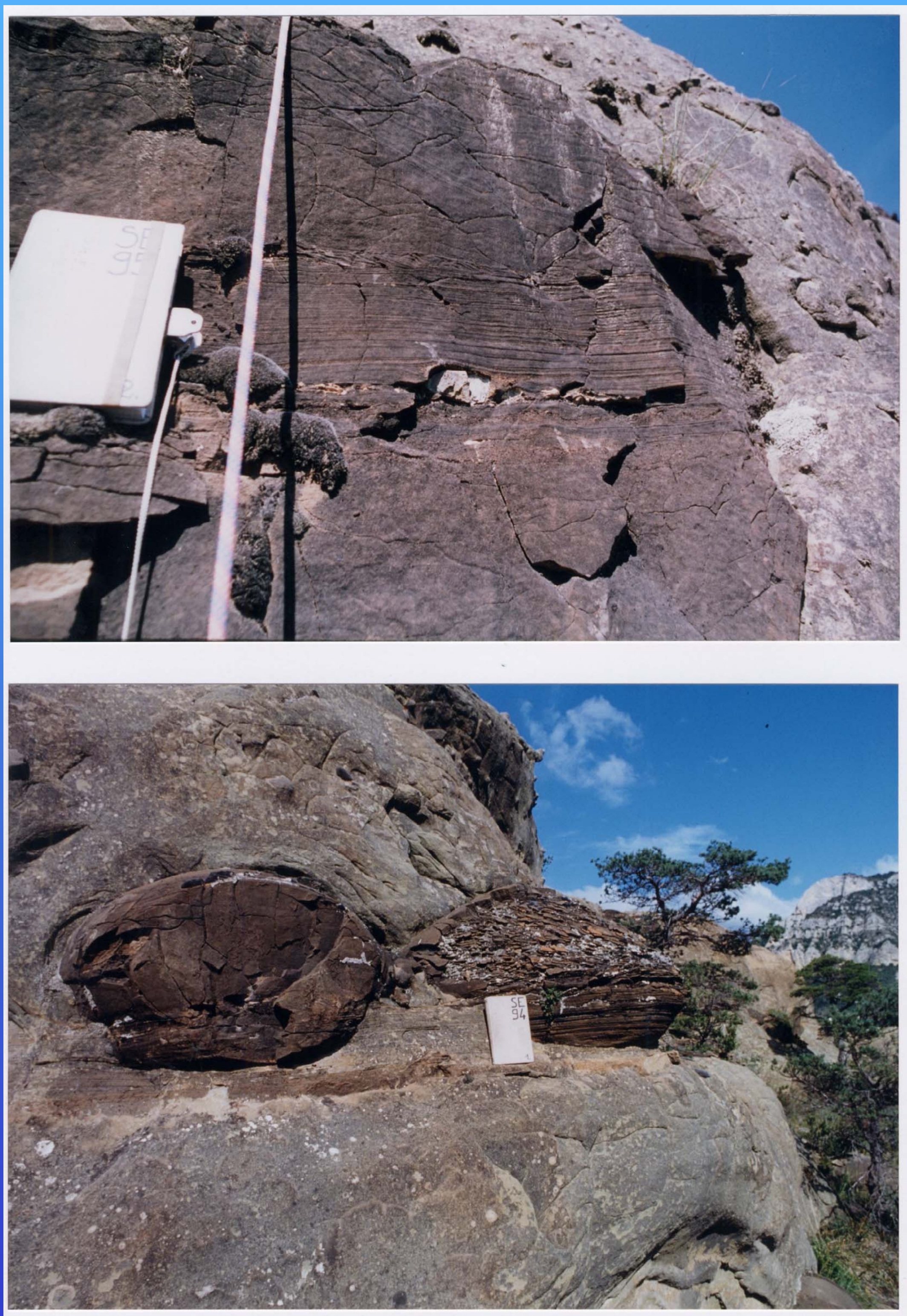
Diagenetic nodules (doggers) crossing parallel laminations



STOP 7: SERRE D'AUTRUY

WALKING FROM THE INJECTITE NETWORK

TO THEIR MASSIVE FEEDER

The northern bank of the T3 Rosans channel complex

We go and stop strictly on the boundary between the injectite (sill-dominated) network and the channel massive infilling.

Internal organization of the channel complex

We are going to walk down the Rosans sandbody. The beds at the top, from which we shall start, show familiar features, i.e. Bouma's Tb-c at the top of a massive sand which could be interpreted as Ta . Going down the section shows less and less well expressed laminations, and very rarely the rippled interval corresponding to Bouma's Tc.

The Rosans sandbody actually consists of the stacking of a number of episodes. Individual events can be distinguished where erosions at the base ise expressed or when an interval of parallel lamînation or ripples appears at the top. Several exposures show the lateral transition from a well defined erosion to an apparent continuity along the basal contact of a bed.

Geometric relations beetween the injections and their feeder

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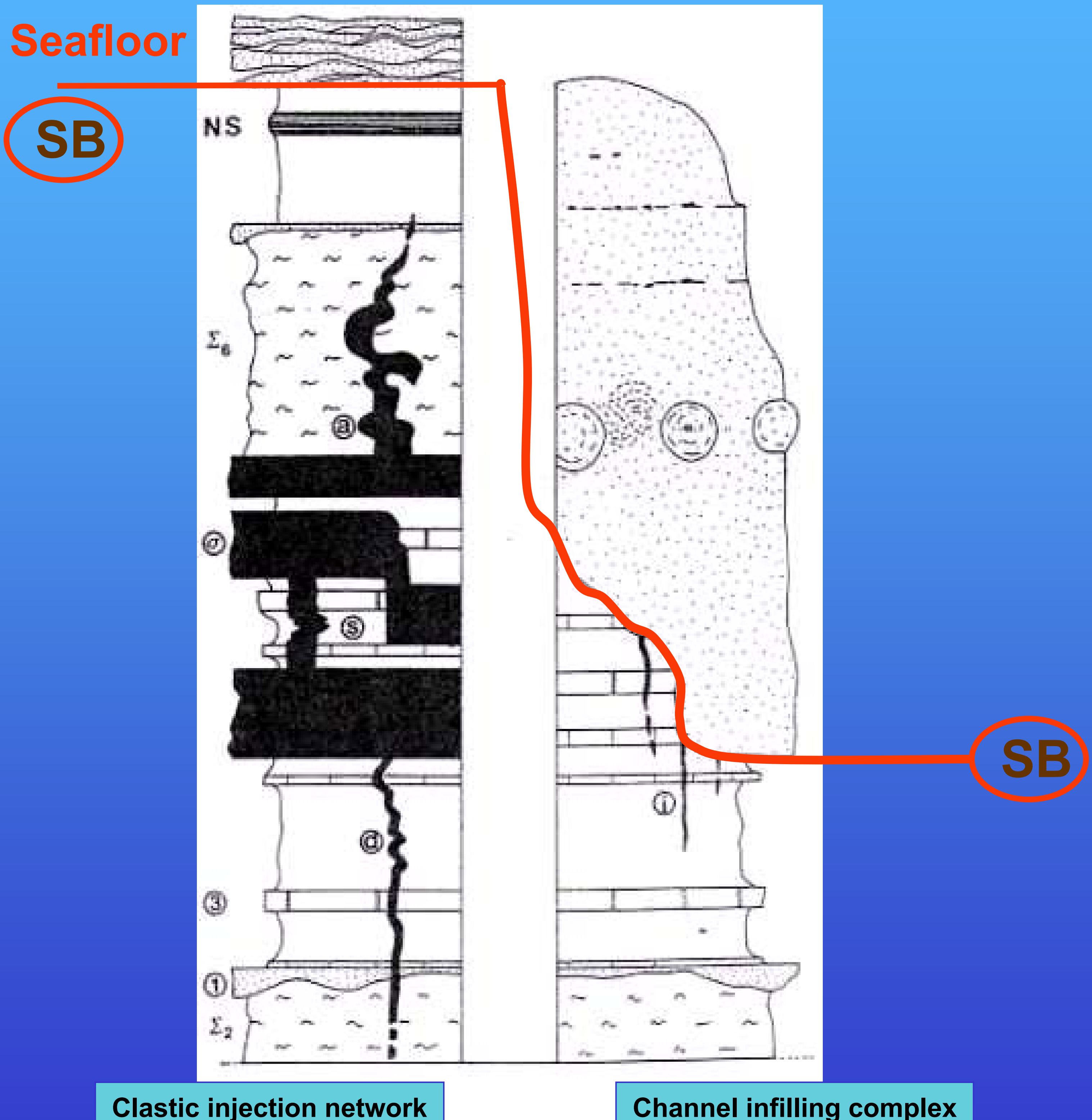


Clastic injection network

Channel infilling complex

Individual massive sand beds, i.e. beds which can be traced laterally without displaying any internal discontinuity, vary in thickness between 0.5 and 4 m in most cases - The thickest single event beds make the lower part of the outcrop and can reach a thickness of 30 m.

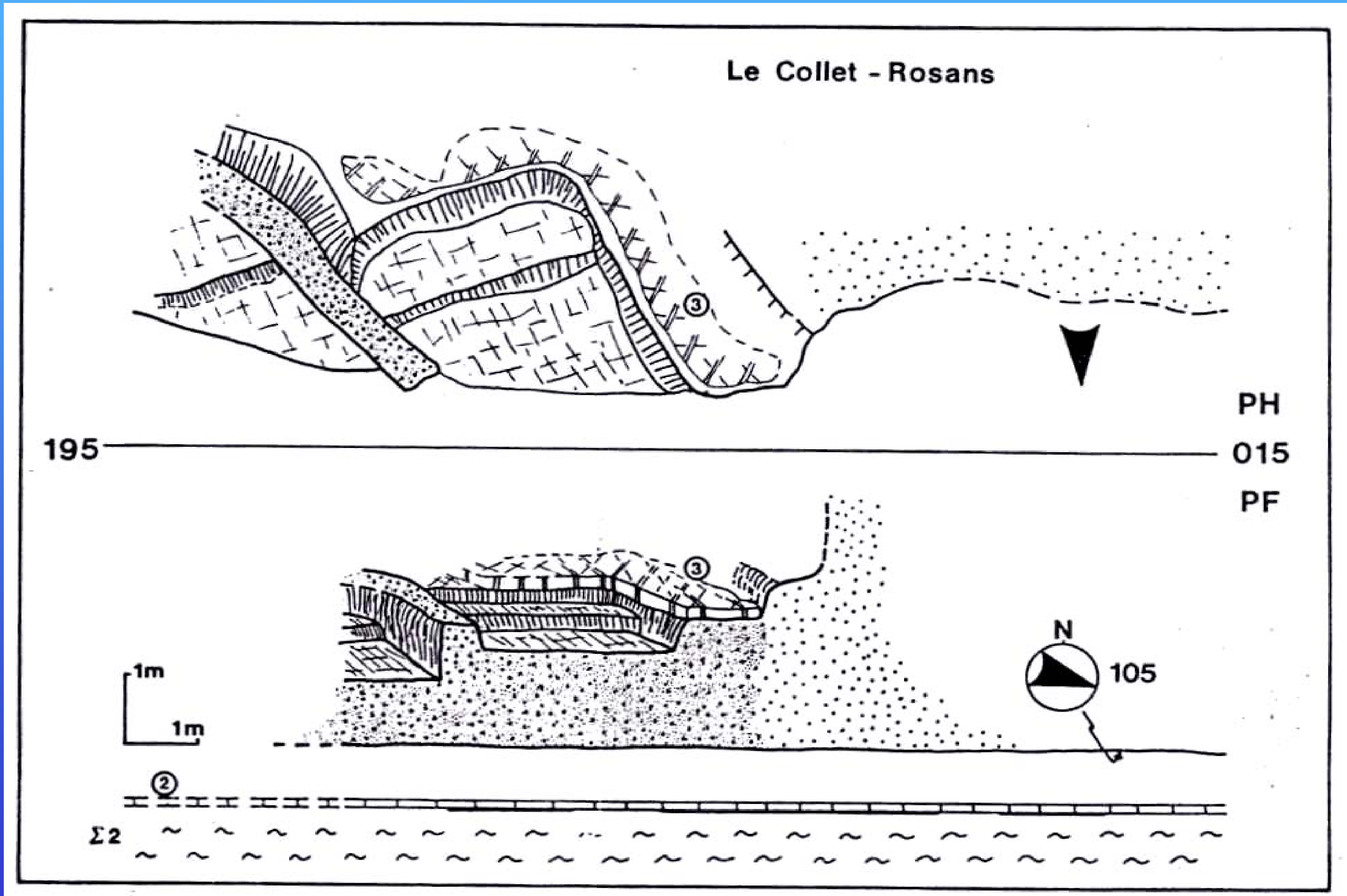
Shale clasts are frequent and often mark the limit between the really massive part of the bed and the upper interval of faint. Diagenetic carbonates usually develop around these shale clasts, making the big spherical doggers which are the most conspicuous feature in the sands. The sphericity is interpreted to reflect an isotrope of permeability in the sandstones.

Megabed, turbidites and flow mechanics

The structure of the beds, as described above, shows many common characters with normal turbidites. The presence of shale clasts is at, or close to the limit between the massive lower part and the laminated upper part of the beds. Such beds can be interpreted in a way much similar to classical turbidites. The main difference lies in their very homogeneous, single grain size character (except a very subtle vertical evolution).

The megabed which appears to result from one single event is less readily understandable, at least as far as the precise depositional mechanism is concerned. As the thin sections cut from various parts of the bed show, massiveness and grain size homogeneity are more than a feeling. It could result either from a large volume flow deposited “en masse” when arriving at the widening point of the valley, or from a sustained flow, the sand being homogenised later by fluidisation.

An other example of relation between channel and injections



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Stop 1-8
Aigret Eastern part
Synthesis

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Sandy levee
or overbank deposits

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STOP 8: LES AIGRETS HILL
EASTERN PART

THE DISTAL PART OF
THE T3 CHANNEL COMPLEX

Walking up the hill at Les Aigrets will take us through the Grand Slump, which may be totally unnoticed to people who have not been informed: the slump is almost undisturbed here. The small turbidite that develops at the top, and is part of the same event, is also unconspicuous and could well be mistaken for an isolated medium-bedded turbidite. A more convincing view of the slump is exposed just across the river.

Les Aigrets channel

The sandstone layer which makes the top of the hill is the distal part of the massive sand seen previously. The thickness here decreases to about 4 m, the distance to Serre d'Autruy being about 2 km. Directional indications (flute casts and so on) at the base of the bed indicate that the flows were coming from the NW, i.e. from Serre d'Autruy. The sandbody here comprises at least two distinct events. The limits of the channel can be seen on the northern and southern sides of the hill: in both cases, the channel passes to sills and dykes.

Lateral facies: "levees" of massive sands

A specific sandy facies develops on the southern side (southern bank of the channel) and consists of decimetric beds with parallel laminations, continuous at the contact with the

channel and then discontinuous. These medium-bedded sands are interpreted to result from an overflow from the channel, i.e. to play the role of levees, during a by-pass phase ?
Similar beds are laterally associated with the channel sands in other places of the basin. They can be traced up to a few hundred metres from their parent channel. Where associated with sills, their lateral extension away from the channel is less than that of the sills.

Mapping of the reservoir

The sandbody can be traced over several km upslope. To the West of the village of St André, i.e. 1 km from the cliff, it is 40 m thick and about 500 m wide: both banks can be seen at the outcrop. A map of the entire system is shown on picture

Controls on reservoir deposition

The beautiful thick reservoir of St André have one way or another managed to travel as far as 40 km away from the shelf that sourced them. Why are they so thin at Aigrets Hill, a mere 2 km from the Serre d'Autruy outcrop? We can see the two margins of the channel, it is not a matter of lateral position; the sands in the other, northern branch of the syncline also die out in a matter of a few km. So what?

The answer appears to lie in the structure: for some reason, St André acts as a sand trap. Two effects can be suspected:

- in the most classic interpretation, turbidity currents start depositing their sediment load when a change in slope produces a hydraulic jump. The paleotopography here, with the presence of the Vaucluse anticline splitting the syncline into two branches could play that role, and probably does.

- another effect, also due to the presence of the Vaucluse anticline, could be the widening of the syncline at this branching point: mapping shows that the sandbody is about 500 m wide just west of St André, the widens to its maximum size at the level of the cliff at Serre d'Autruy, and the two distal branches are narrow again (a few hundred metres). The widening of the topography must have decelerated the currents (not necessarily killed them, sustained currents could be responsable for some of the laminated beds observed), thus inducing sand deposition.

Notes

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SILLS, DYKES AND FEEDER

A block diagram shows the pre-compactional relationships between the injected sands and their feeder body. The relationships observed between the injected sands and their feeder sandbodies in the Vocontian basin indicate that they were injected during the deposition of the sands, in most cases downwards or laterally from their feeder.

Sills and dykes deserve a special mention here: they are best developed in St André, i.e. where the massive sandbody widens and dies out. They appear in a way to play a role much similar to lobes in flysch contexts, by dissipating the kinetic energy of the flows.

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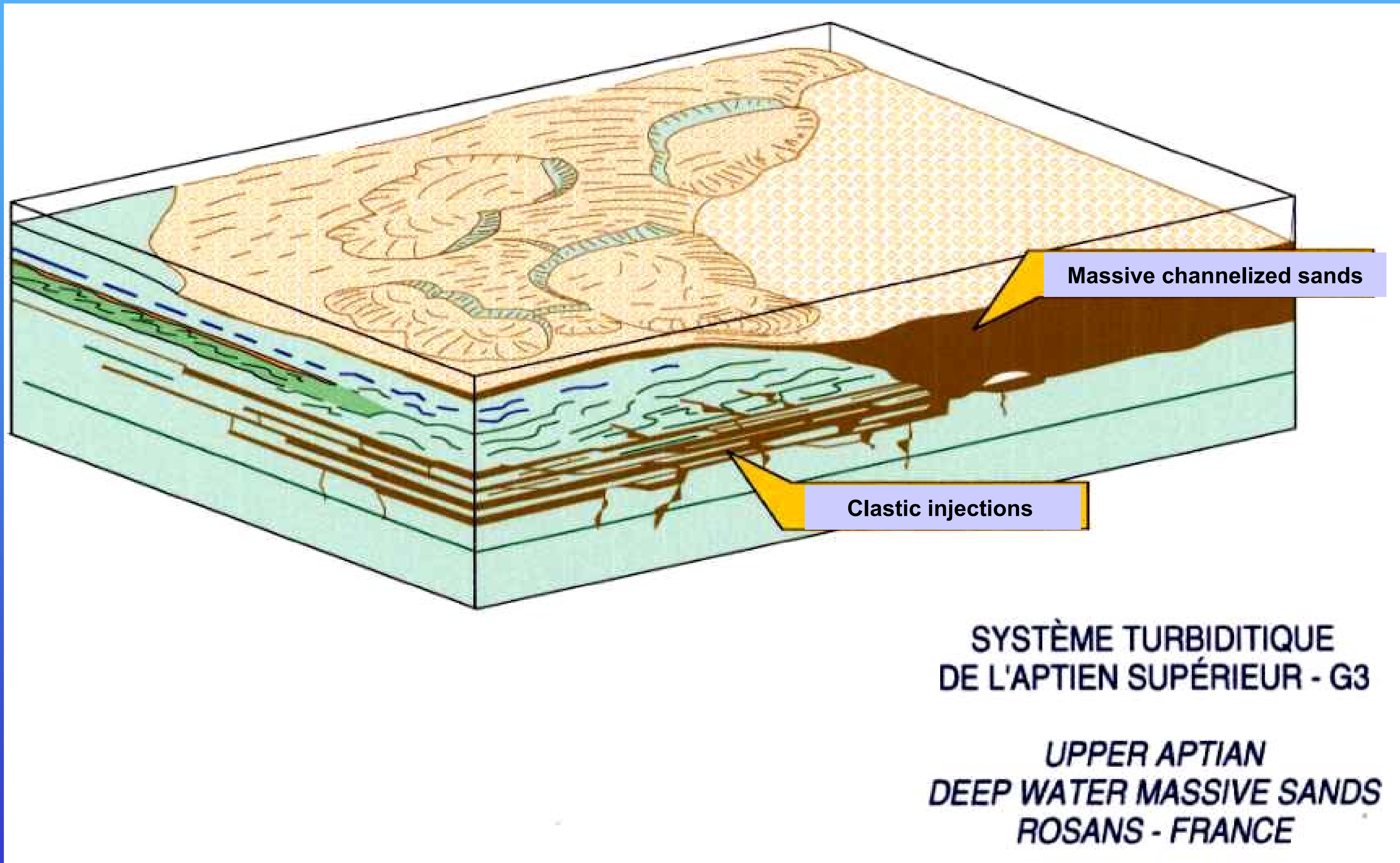
Rosans sandstones: a geometric model



« Classical » slump in Goguel (1938) from Gras (1835)

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Stop 1-9

Montclus

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From *in situ* sedimentation to debris flow deposits

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Vocontian Clastic Injections

September 2001

STOP 9: MONTCLUS

HAUTERIVIAN AND BARREMIAN CARBONATE

SLUMPS AND DEBRIS FLOW DEPOSITS

Numerous pretty slump or debris flow deposits crop along the road. It is the most spectacular section in the Vocontian area showing mobilised carbonate deposits

Notes

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DAY 1:
BEVONS AREA

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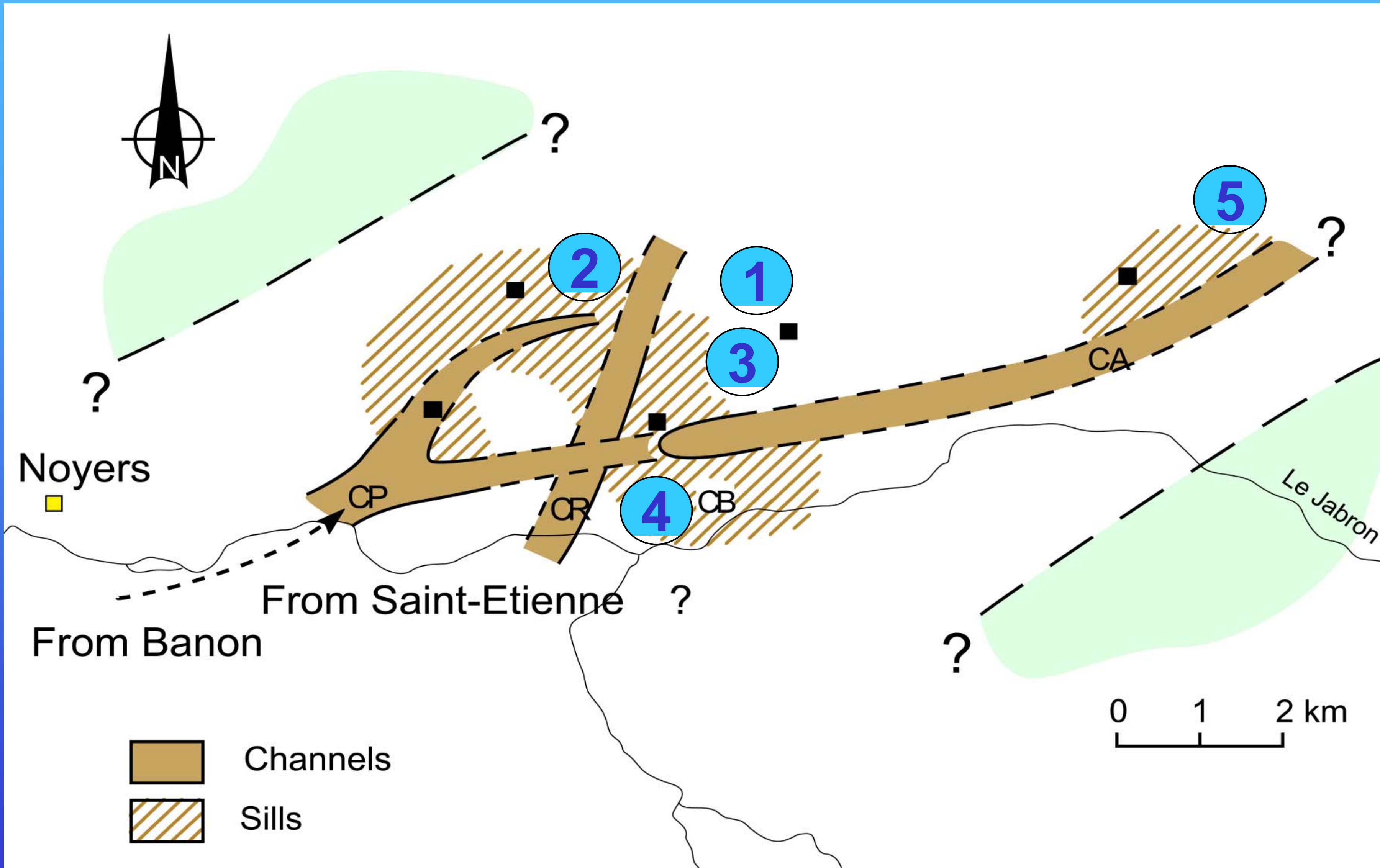
Day 2 Stops


N
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Noyers
CP
CR
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Le Jabron
From Saint-Etienne
From Banon
0
1
2 km
Channels
Sills

Saturday 15th September 2001:

VOCONTIAN CLASTIC INJECTION EXAMPLES:

DYKE AND SILLS NETWORK,

DYKE PATTERN.

RELATION WITH THE FEEDER AND THE BEDROCK

Other topics: massive sands, channel infilling, compaction, topographic control on sand deposition, syndepositional deformation and faults, early fracturing.

Bevons - General setting

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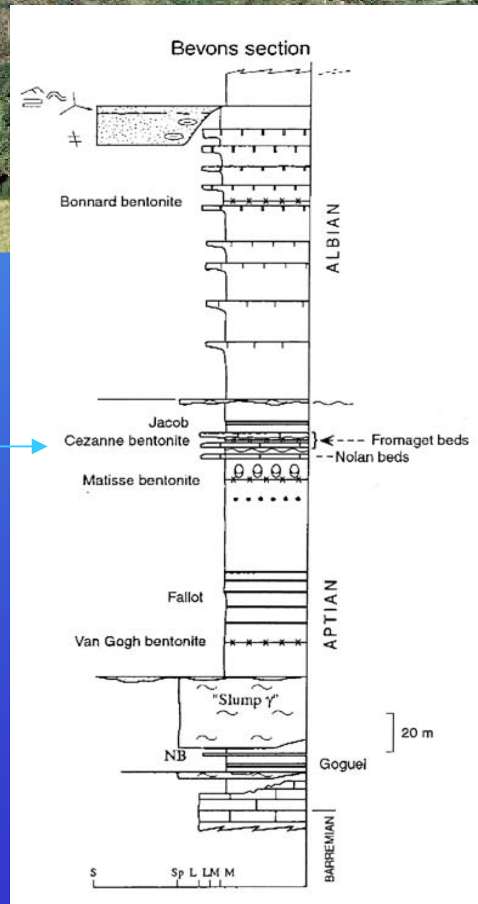

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Stop 2-1

Bevons school

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Litho-stratigraphical setting

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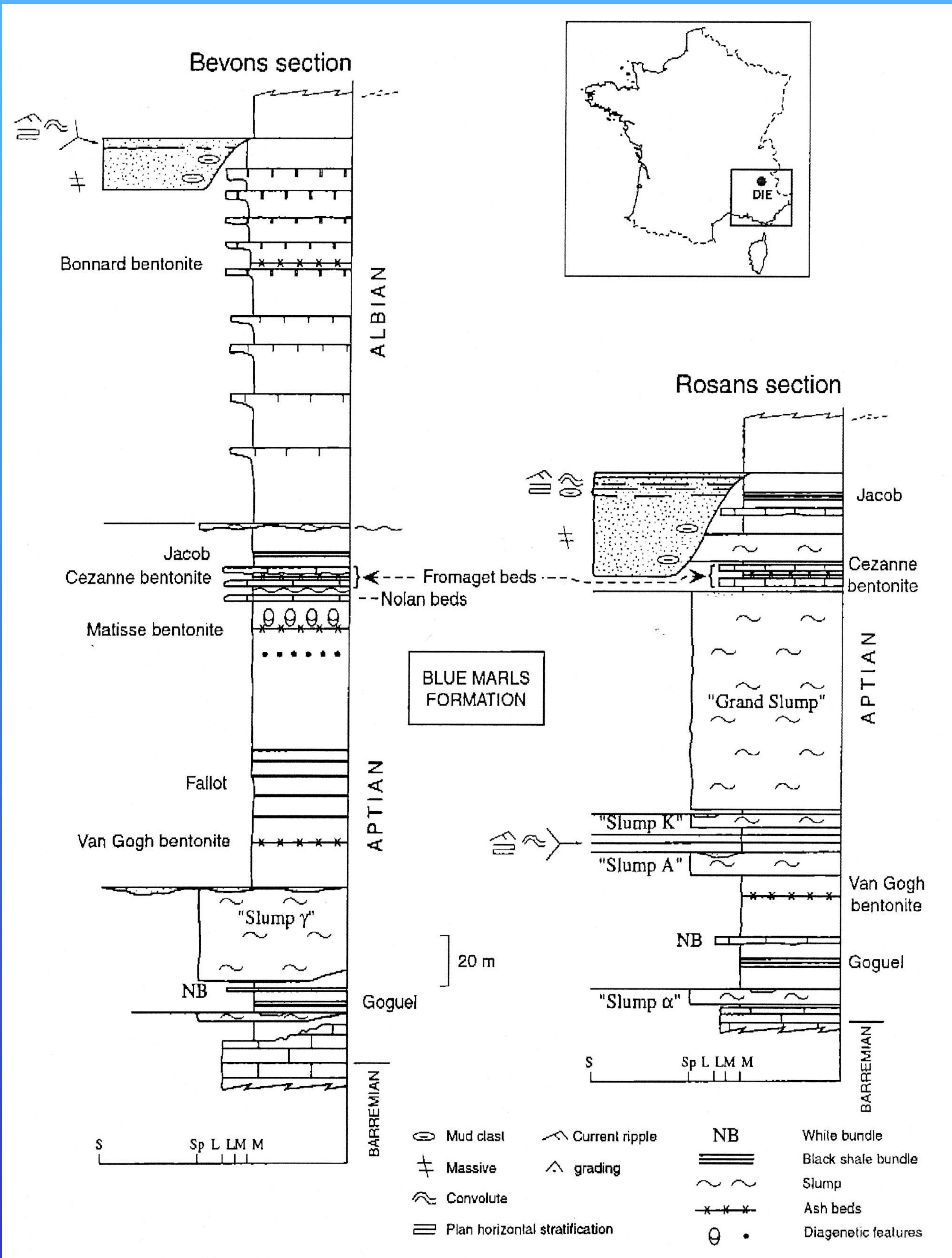

STOP 1: BEVONS - SCHOOL

GENERAL STRUCTURE AND STRATIGRAPHY OF THE AREA

Since the end of the last century, Bevons is a reference section for the Aptian and Albian series North of the Montagne de Lure. The so-called “Blue Marls” formation consists of a 300 metre thick marly unit capped by 20 metres of sand bodies sitting on top of the hill. The sandstone layers are of Uppermost Albian (Vraconian) age (G8), three different lenses have been identified, CB, CP (same channelled system or C2) and CR (Channel system C1).

No other sand body is present in the Aptian-Albian series! And downward the nearest sandstone body is fluviatile without glauconite, Triassic-age and located downward up to 10 kilometres!

Mapping of turbiditic channels in Bevons area

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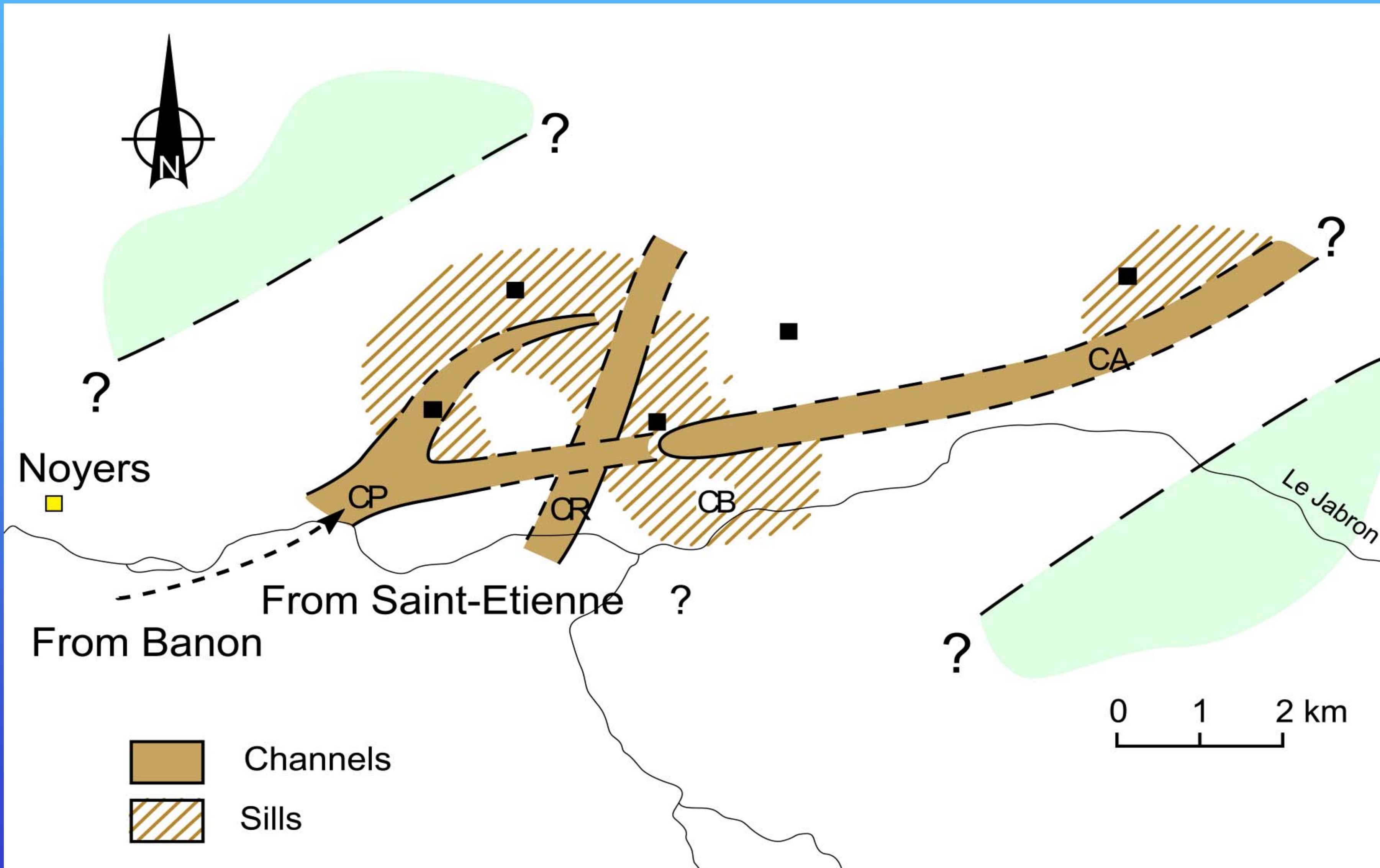

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Stop 2-2

Le Puy Hill

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Le Puy Albian sandstones: the Cp complex

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STOP 2: BEVONS – LE PUY HILL

GEOMETRY OF THE CLASTIC SILLS AND DYKES NETWORK

The Bevons sandbodies

The first observation on this outcrop is the geometric continuity of each sand body, whatever its complexity. The detailed geometrical analysis will then allow defining characteristic shapes.

In order to better understand the complexity of this outcrop, a quick drawing is required. It will help along with the analysis and progressive characterization of the different geometry of the sandy bodies.

Channels

These sandy bodies correspond partly to true erosive channels; in fact we can identify flanks and sole marks. The mapping of the channels and the current indications (sole marks) show a slightly sinuous path on the slope of the Vocontian southern palaeo-margin. The sand comes from the southern shallow water Montagne de Lure platforms: the clastic influx bypasses the Banon paleo-valley like the upper Albian megaturbidites (G7).

Your drawing of Le Puy hill (1)

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Your drawing of Le Puy hill (2)

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DE PARIS

Injectites

The bedding is clearly underlined by the abundant limestone beds. Some of the sandy bodies are oblique to the bedded formation. Therefore they are clearly injectites that will be described in detail during this stop.

The sandstone dykes

The dykes are an easily recognisable form of injection. Some of them are several hundred metres long and have tens of centimetres thick walls. Their thickness varies from few millimetres up to one metre.

The Northern outcrop of Le Puy Hill presents a dense and complex network of sandstone dykes. The dykes can be followed directly here over more than 150 m from the Uppermost Albian sand bodies down into the Upper Aptian marls (Gargasian). Further investigations have shown that this network extends much further down, cutting through the entire Aptian section where no clastic event has been recognised. Therefore the depth of penetration of these dykes is roughly 300 m (not decompacted).

The sandy infill of the dyke records the post-injection compaction. The primary geometry has been restored just after injection (“decompaction”); it demonstrates (i) a different fracturing pattern between the marls and the limestones and (ii) a differential compaction rate according to the vertical distance from the sandy channels.

The diagenetic evolution observed in the primary clastic infill and in the walls of the dykes (limy concretions) demonstrates the key role played by the injections in drainage.

The clastic sills

The present morphology of Le Puy Hill is related to sandy turbiditic channels but also to clastic sills (up to 10 metres thick) which are directly associated with the dykes in geometrical continuity.

Step-shaped geometries linked with change in their stratigraphic positions characterise the sills. However, smoother changes are also observed, as the sills gently penetrate the host formation at very low angles.

No primary current marks (sole marks) are to be found at the bottom of the sills; similarly no sedimentary structures (laminae, ripple…) have been observed. The walls are very flat and regular, with localised frondescent casts of diagenetic origin near the major steps. The sandy infilling is homogeneous except at the connection with the dykes where between one to five centimetres shaly intraclasts are present.

The clastic sills are restricted to the upper 35 metres (compacted) of the section. The lowest stratigraphic level is marked by the metric sandy “bed”, bed 31, connected with overlying and underlying dykes. Located near the sequence boundary A3/A4 (Friès, 1987), it corresponds precisely to a centimetric yellow-orange clayey horizon interpreted as an ash band (Bonnard level, called after the French painter). Below the channel CP bed 31 becomes a massive plurimetric sandbody with no sole marks but a few frondescent casts and connected to clastic dykes.

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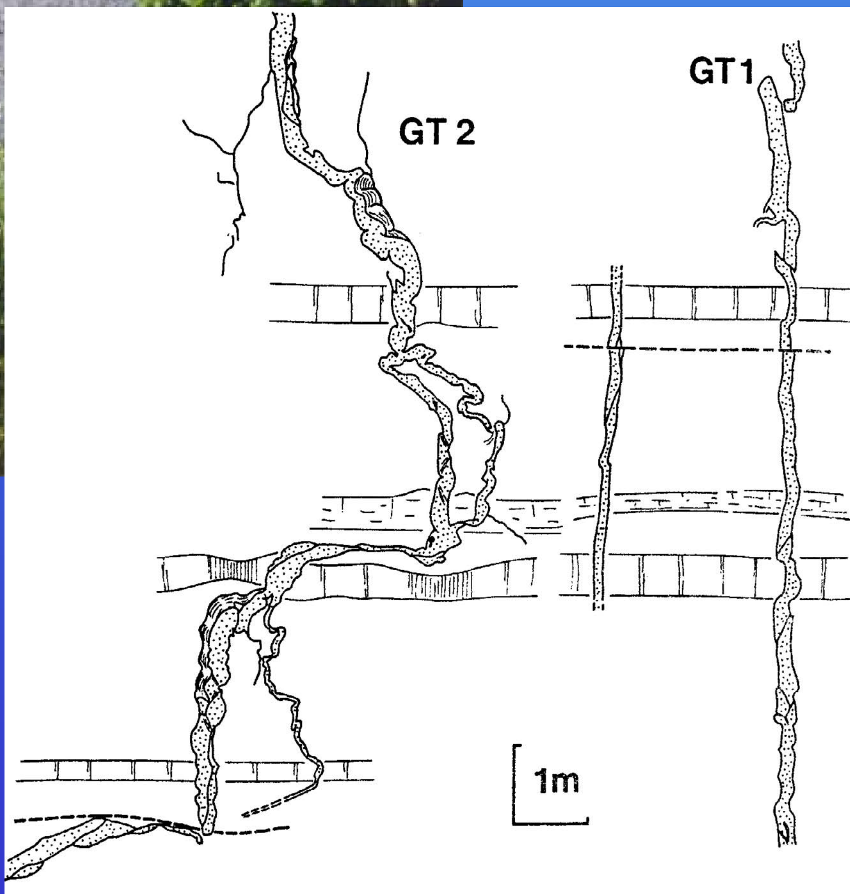

Vocontian Clastic Injections

September 2001

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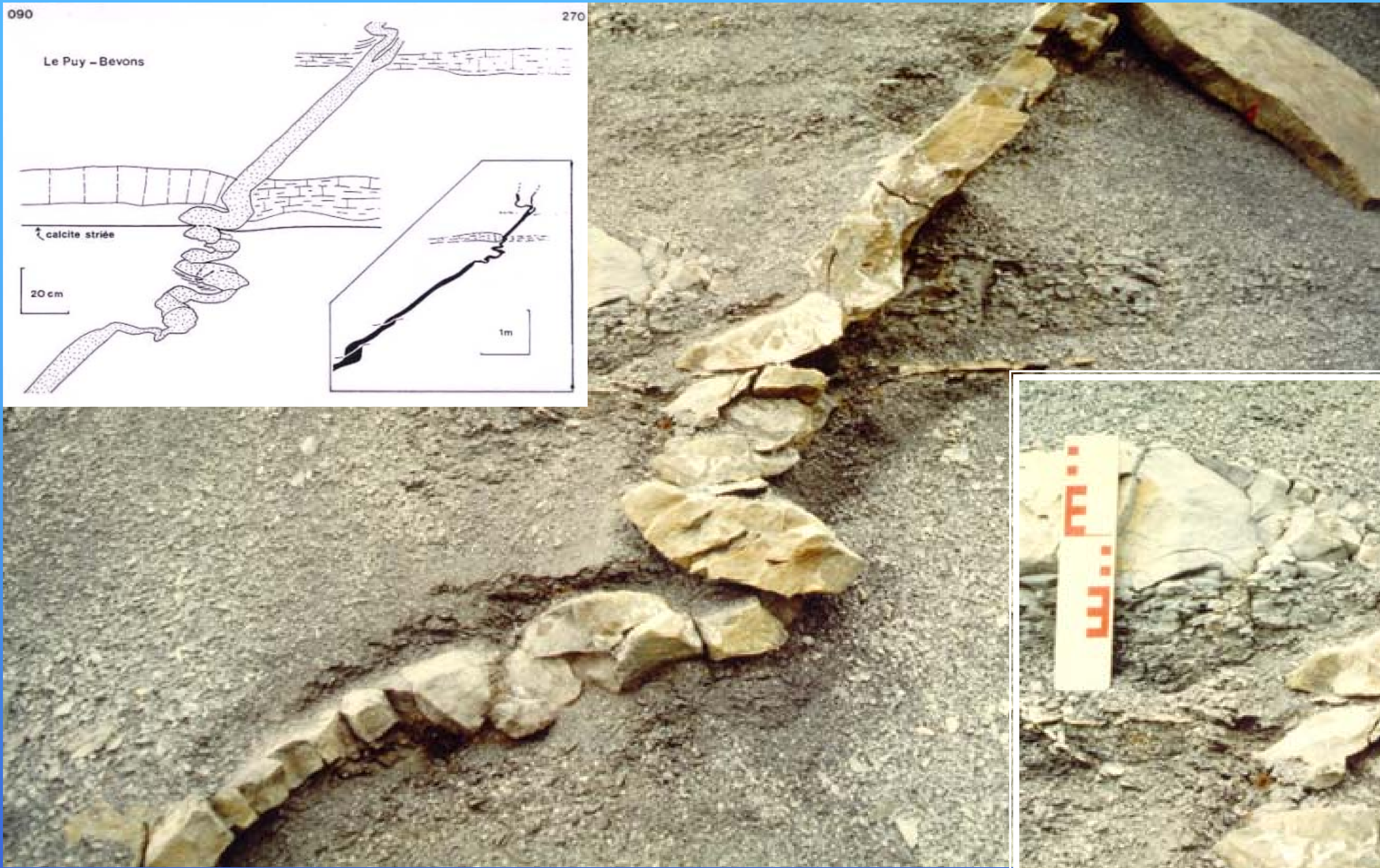

Differential compaction

Relation between sills and dykes

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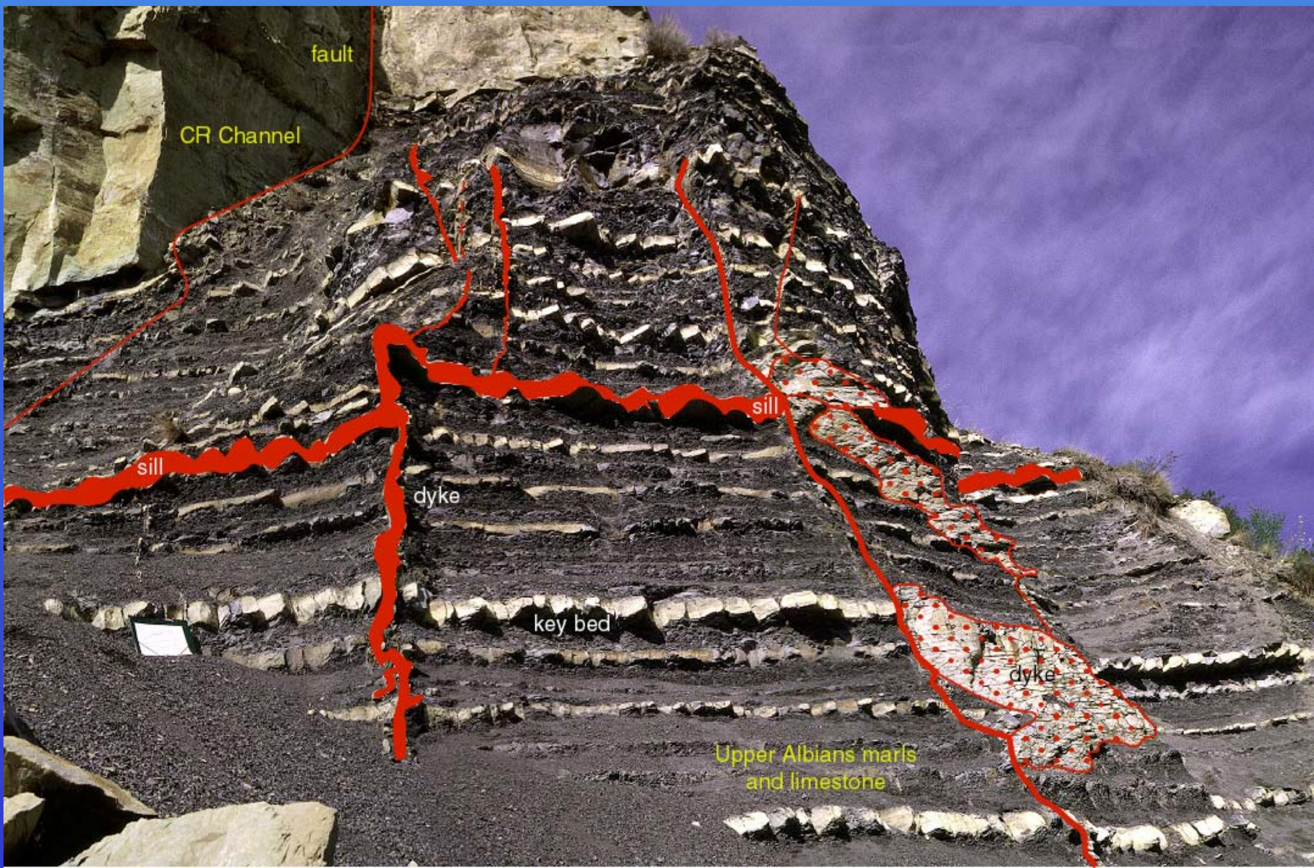

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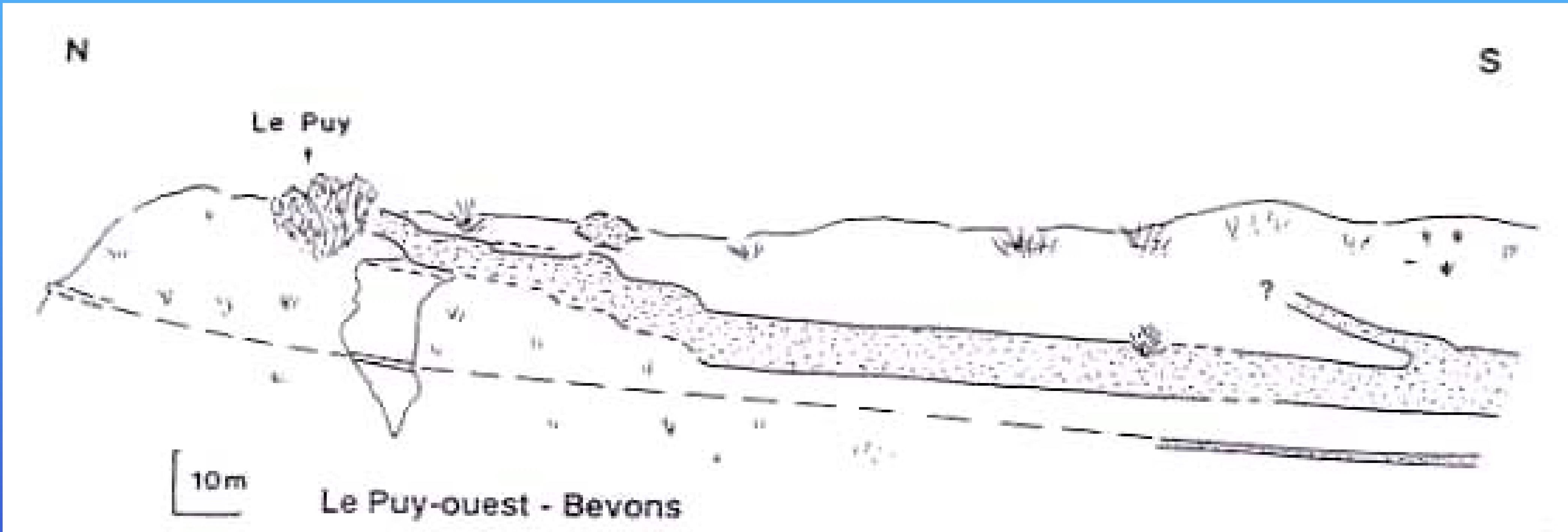
Large scale sills in Bevons area



Bevons Cp complex: Injections developped in the channel margin

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The channel deposition story

The channels are several hundred meters to one kilometer wide; they are filled in by one major event (channel CB) or by amalgamated events (channels CP and CR). The massive sandy bodies result from high density turbiditic (*sensu* Lowe) events.

The C1 Bevons channel

This channel is the first observed during this stop. The measured sole marks (flute and grove casts) at the bottom of this body are coherent and indicate a flow towards N010. This sub-meridian channel has been continuously mapped from Les Rouines (southern side of Le Puy Hill) to the northern side outcrop. North of Les Rouines it is locally absent on several hundred square metres.
Two main events infill the channel; they are five to six metre thick each. However, from section to section the infilling varies rapidly. The total thickness reachs 15 metres as for example on the northern outcrop of Le Puy Hill.

The C2 Bevons channel

This western channel can be traced over three kilometers from Barneche to the top of Le Puy Hill. Its ENE-WSW axis is parallel to the paleocurrent indications. On the northern side of Le Puy Hill its width is only 50 metres, whereas five hundred metres to the South, it is much larger – about 200 metres. **The detailed mapping demonstrate that the channel downwards ending correspond to the northern outcrop of Le Puy Hill; *in situ* sandy facies is replaced by an injected complex of sills and dykes.**

The total thickness of the channel is 15 metres on the Le Puy northern outcrop. At least four different events have been identified in the infill; the lower surface of each event is slightly erosive but it does exhibit any mud drapping. Convolutes and liquefaction figures are locally present on top of the sequences.

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Bevons CP channel margin (close view)

A sandy complex of sills and dykes

The C2 channel is associated with a lot of clastic sills especially on its right flank. The sills present soft steps and mixtures facies (sand and shale). It is also connected on its left bank with a several meter thick sill which is characterized by large frondescent casts. This sill is connected with the major sandstone dykes observed and described previously.

In conclusion the channel CP is geometrically linked with a complex network of clastic sills and dykes. This injections are also lateral or under the channel feeder and the paleo seafloor; moreover no other sandprone body exist in the Aptian-Albian series. Therefore, **the injections are fed *per descensum*.**

Differential compaction

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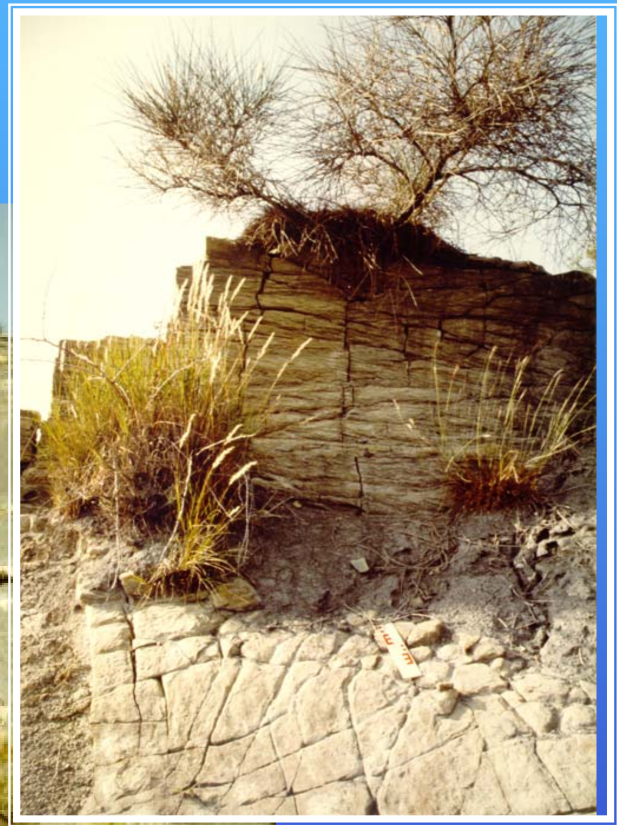

Side view

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Stop 2-3

Vieux Bevons Hill

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Compactional features

STOP 3: VIEUX-BEVONS

THE DYKE NETWORK

A example of dykes network

The dykes are organised in networks and they never exactly cross each other. Their orientation pattern shows major trends parallel to sealed synsedimentary faults of the underlying Barremian-lowermost Aptian limestones. As no throw is associated with the dykes, the fractures are filled with sand and the corresponding walls are only moved aside. However the compaction recorded by the dykes shows a horizontal component; similar results have been documented by Cartwright (Imperial College, London) in his published papers on his work on fractured shales from the North Sea.

Discussion about differential compaction and dyke deformation

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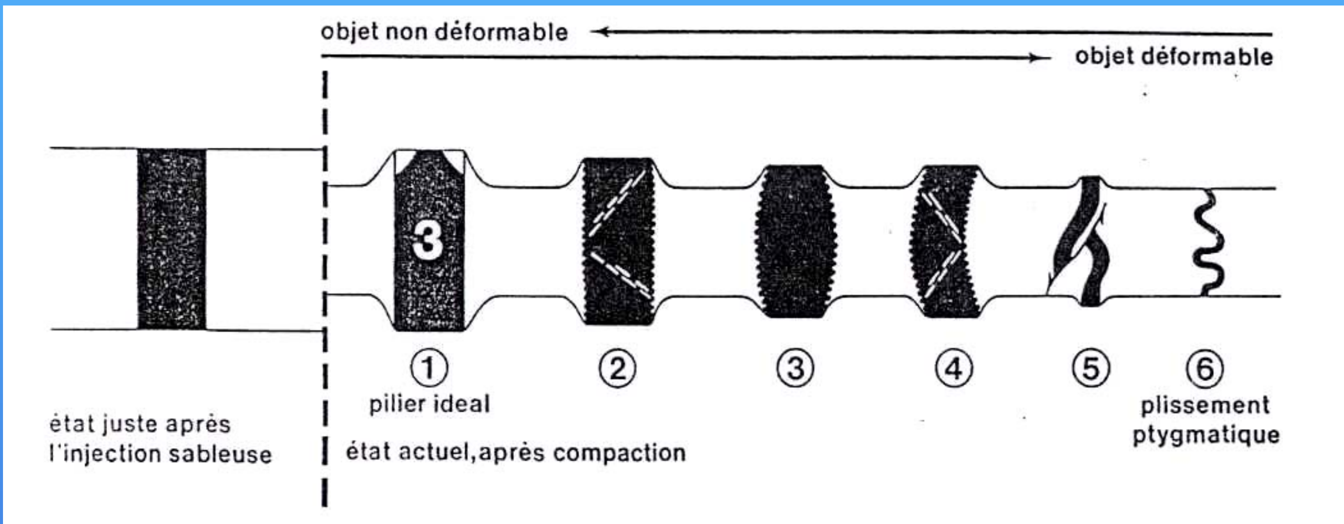
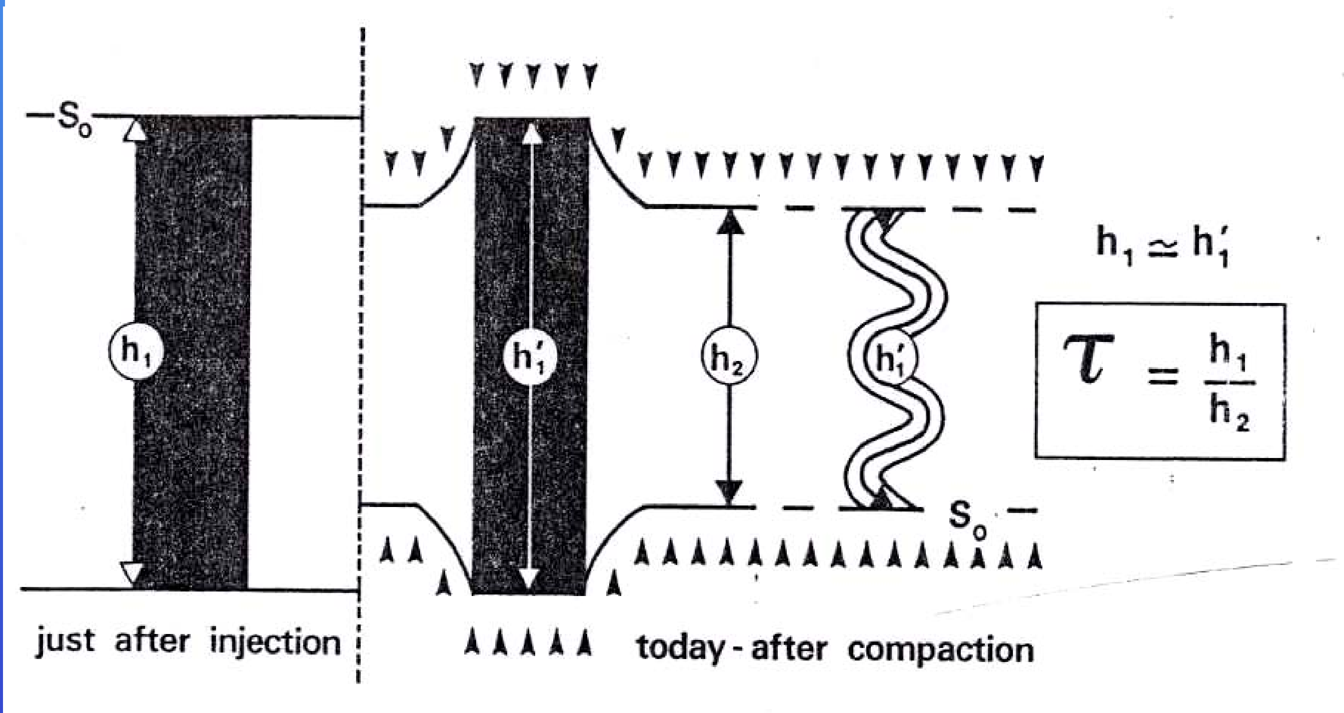
DIFFERENTIAL COMPACTION:
From outcrop to compaction curve





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Stop 2-4

La Baume

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La Baume outcrop

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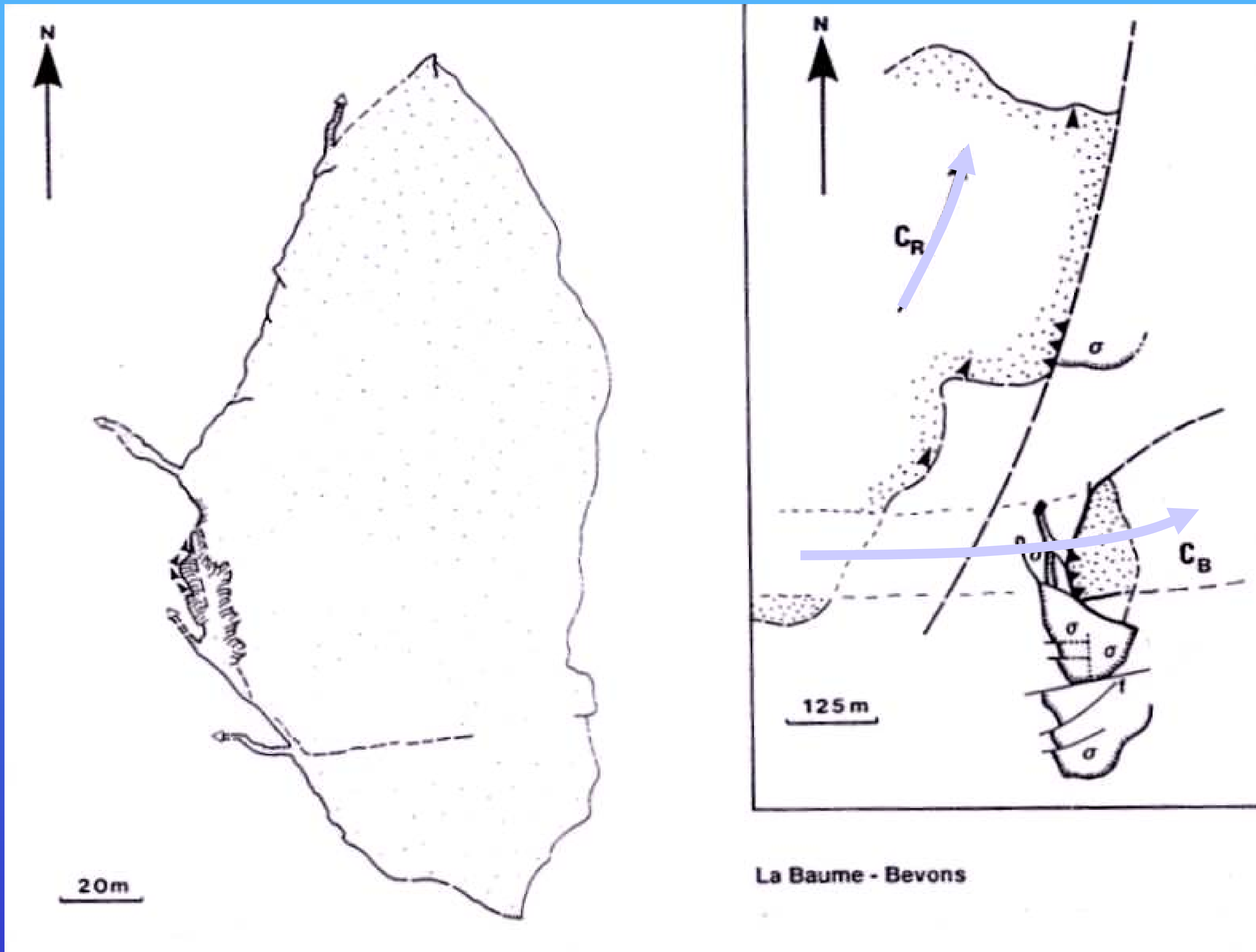

STOP 4: BEVONS
LA BAUME RAVINE

CHANNELS AND INJECTIONS RELATIONS

The C1 Channel

This channel has already been observed on the Northern side of Le Puy Hill. North of Les Rouines, it is absent over several hundred of square metres. The base of the channel infilling exhibits erosional features 1 metre wide and several tens of centimetre deep. Laterally the massive sand body passes into decimetric turbidites. On the western side of La Baume Ravine, the channel flank is almost vertical and 17 m deep. The channel's width reachs here 500 metres.
Several decimetre thick clastic sills are fed laterally from the channel into its flank.

La Baume clastic sills and dykes

The eastern flank of La Baume Ravine is characterized by a complex outcrop including clastic sills, dykes and the CB turbiditic channel.

The CB channel is limited to the southern part of Le Puy Hill. It is filled with a massive structureless sand. Its western flank exposes a spectacular 12 metre deep undercutting. On the corresponding "bottom" surface decimetre long flute casts with specific geometries record peculiar flow dynamics: (i) helicoidal-shaped flute casts that are horizontal and parallel to the palaeo-morphology (ii) vertical, isolated or regrouped flute casts that have fossilized turbulent upwards currents along the undercutting.

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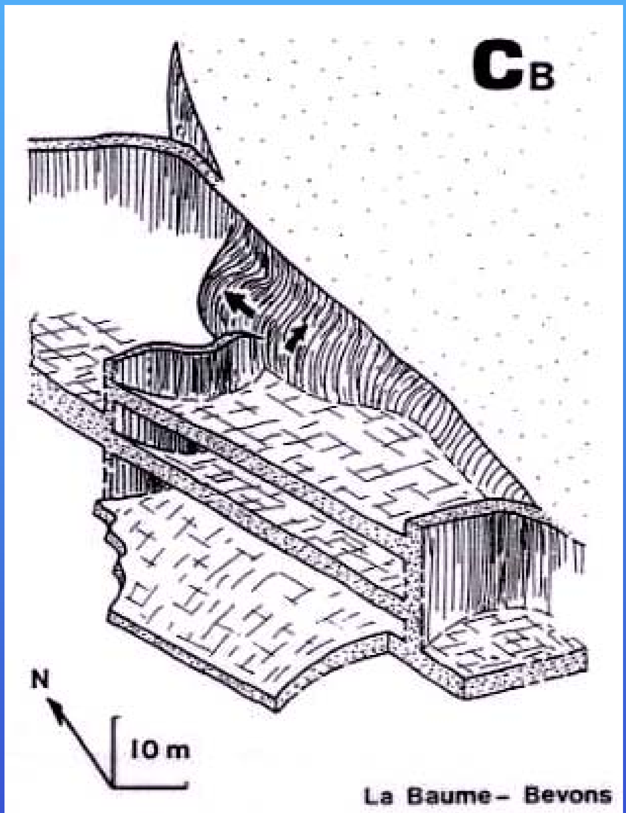
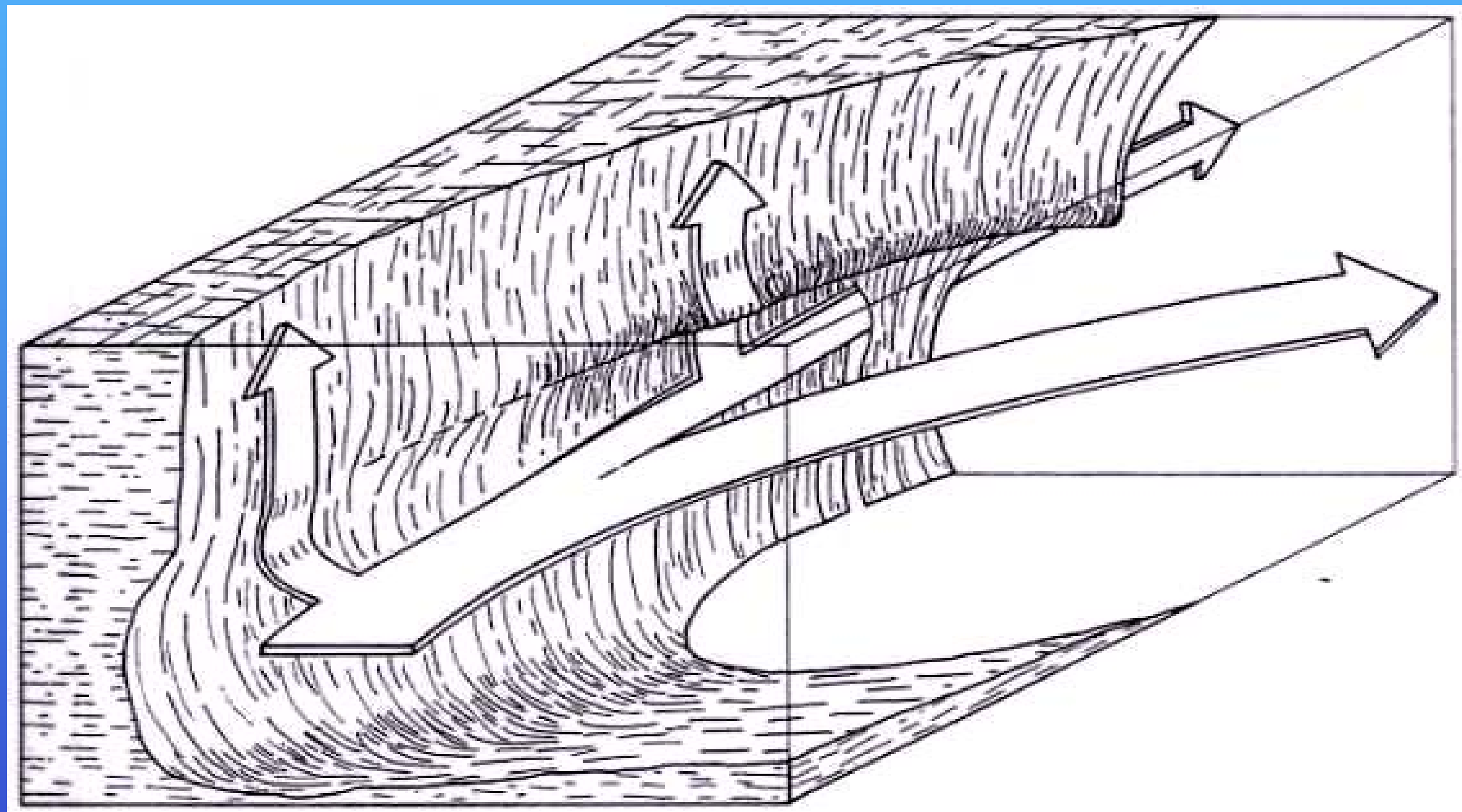
Proposed fracturation and injection mode





The channel feeds four thick (up to 3 metres) main dykes which are injected into the concave bent of the channel. From them, a complex network of thick sills and dykes cut into the marly formation. Locally these injections creates a box-shaped complex network, which is very complicated to decipher.

The outcrop records a major East-West erosion (mega-flute cast) linked with the La Baume main NS fault scarp. The CP and CB channels are genetically linked: the NW CP channel is ending on the northern side of Le Puy Hill, whereas the WNW CB channel erodes deeply into the Albian formation. In this hypothesis, the Pierre Avon sills are also fed by the CB channel.

The channel depth is around 15 metres, and its maximum width is 160 metres.

The Bevons Albian sandstones: a synthesis

The Upper Albian sand bodies of Le Puy Hill are the downslope part of turbiditic system heading 50 kilometres upslope, on the Southern side of the Montagne de Lure, where the channels are cut into very shallow water platform deposits.

The Bevons sand bodies are divided into two categories: true turbiditic channels and clastic sills. Only two main channels are necessary to explain all the outcrops.

- A first erosion cut into the Albian marls (over 15 metres). The channel is filled in with successive massive sand bodies. N010 synsedimentary faults control its depositional pattern. Few injections (sills and dykes) are associated with this first system. However, these injections present a important differential compaction pattern.

- The second system is associated with a complex network of sills and dykes which extends on several kilometres. It erodes deeply the previous system.

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Lunch

Along Jabron river

Provençal **Pic Nic**
with the support of Geneviève and Didier from
"Ferme Auberge Vaujeala"

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Stop 2-5

Pierre-Avon

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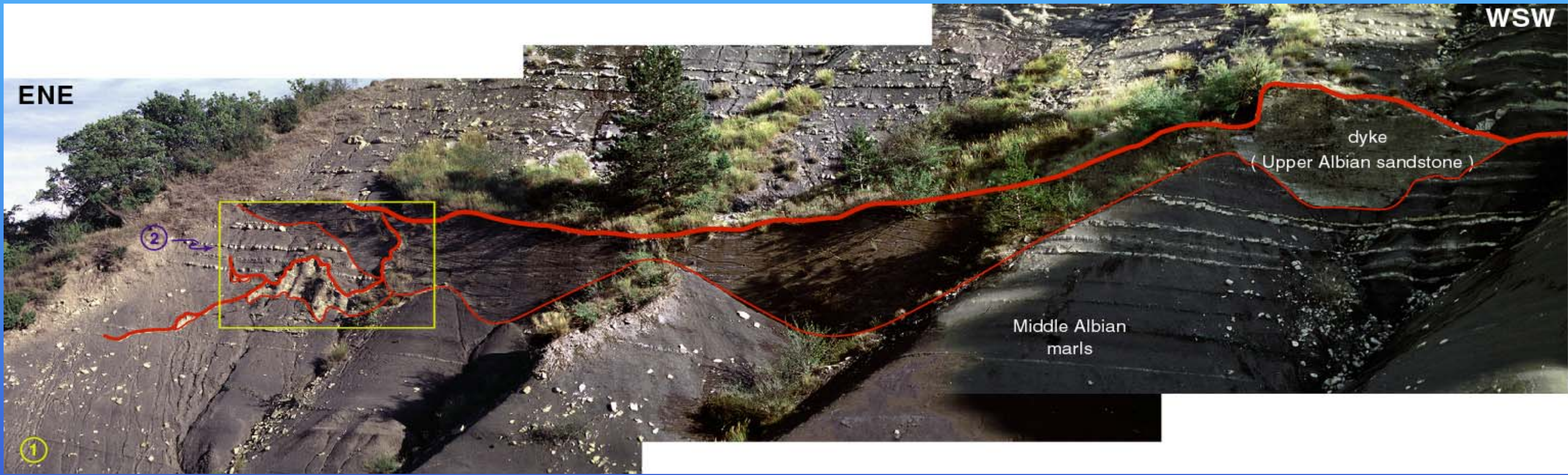
Pierre-Avon dykes network



N.B. These lenticular features aren't channels !

STOP 5: PIERRE-AVON

A DYKE EXAMPLE AND THE DYKE NETWORK

An example of dyke network

The structured “Barremian-Bedoulian” substratum (if the night…)

Relation between fossilised faults or escarpements and dykes network

Mapping of Pierre-Avon dyke network

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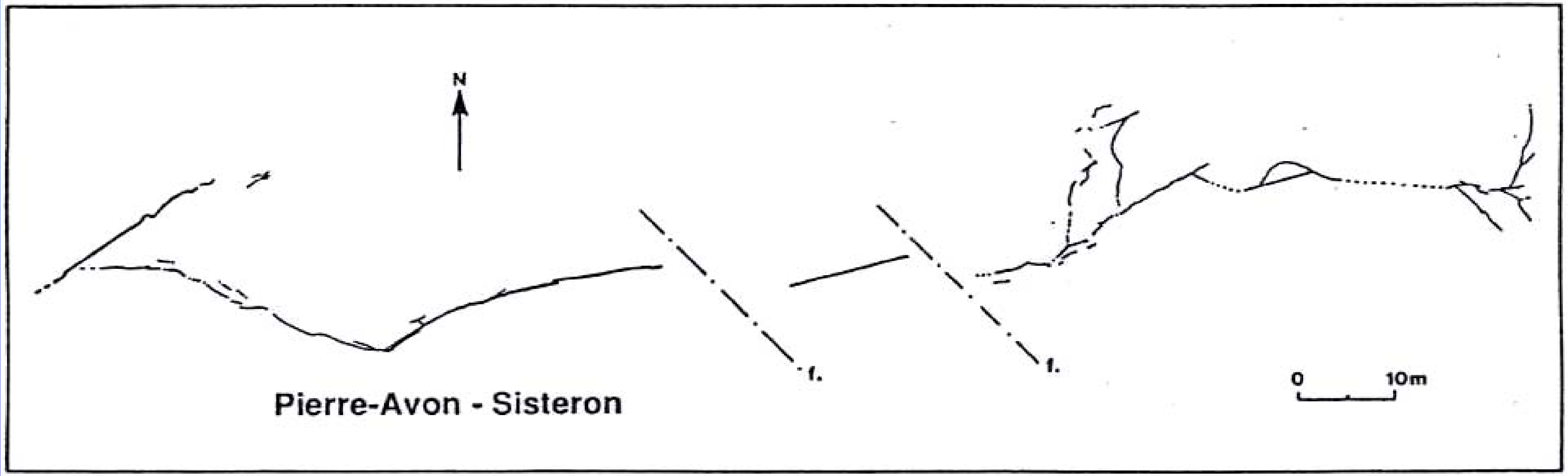

Dyke vertical organisation

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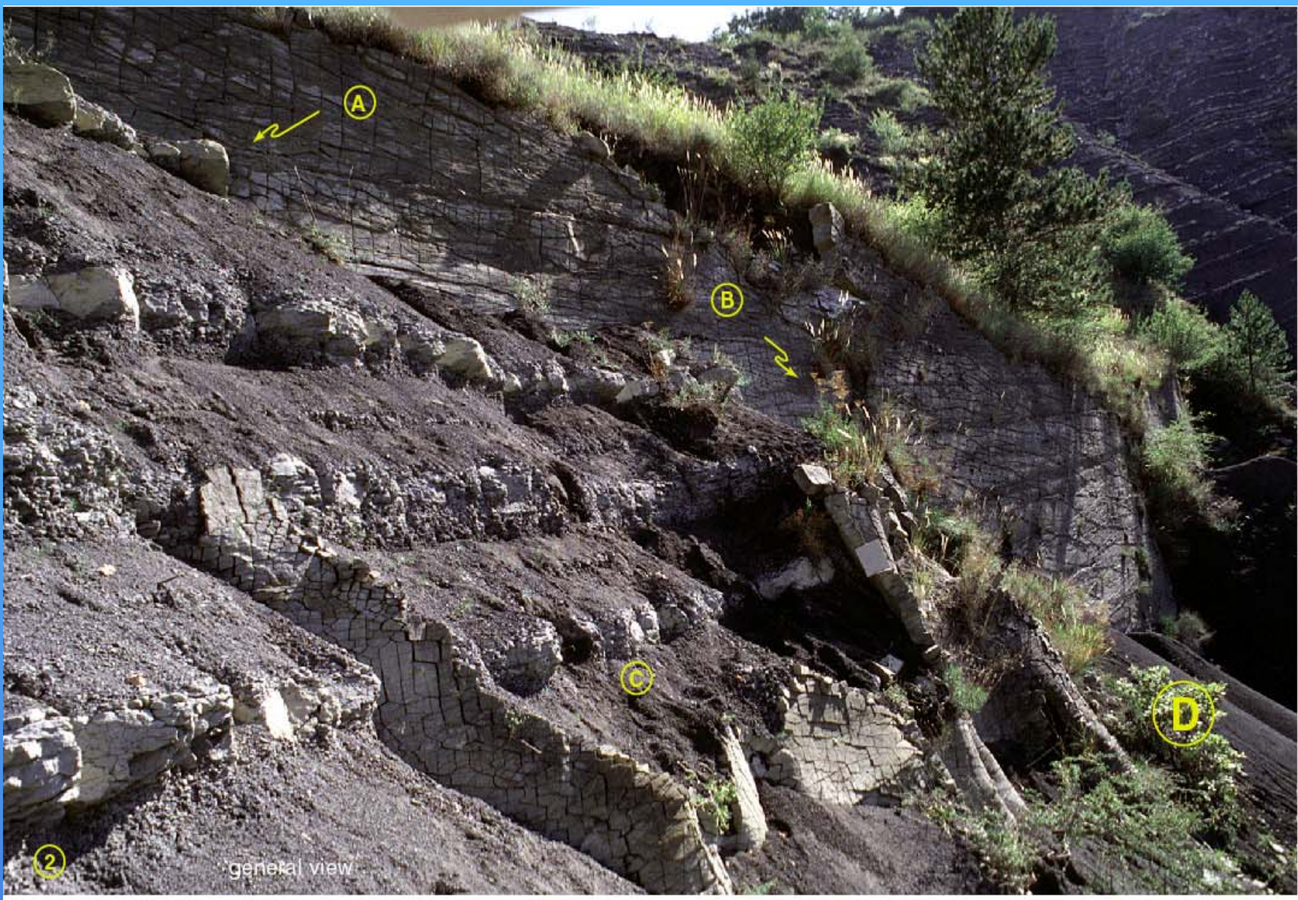

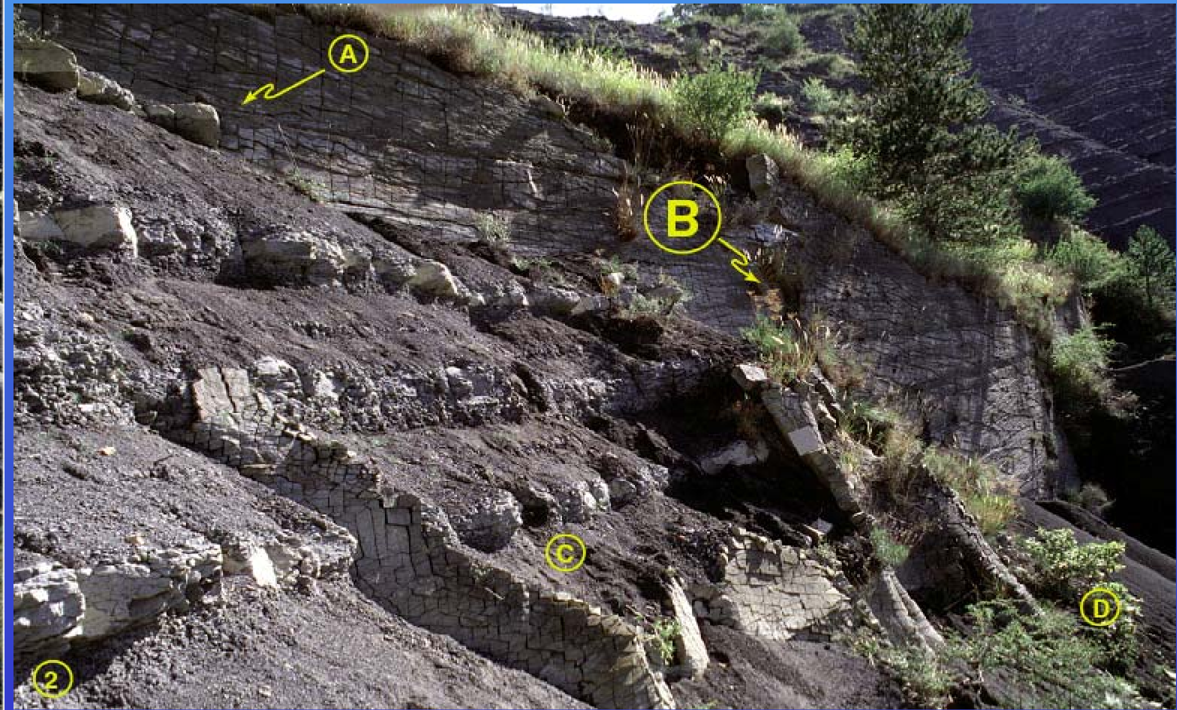

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Relation between injection shape and host lithology

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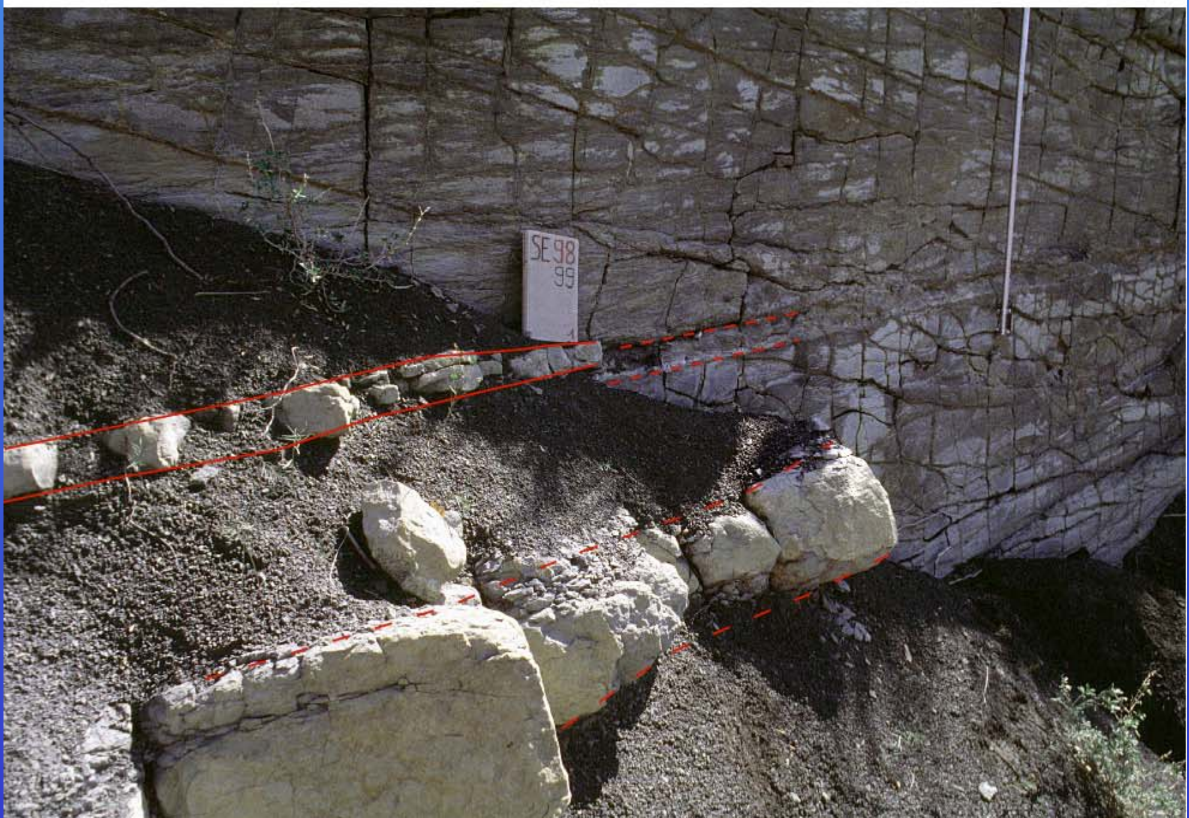

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Perspective view

Map view

Dykes network organisation

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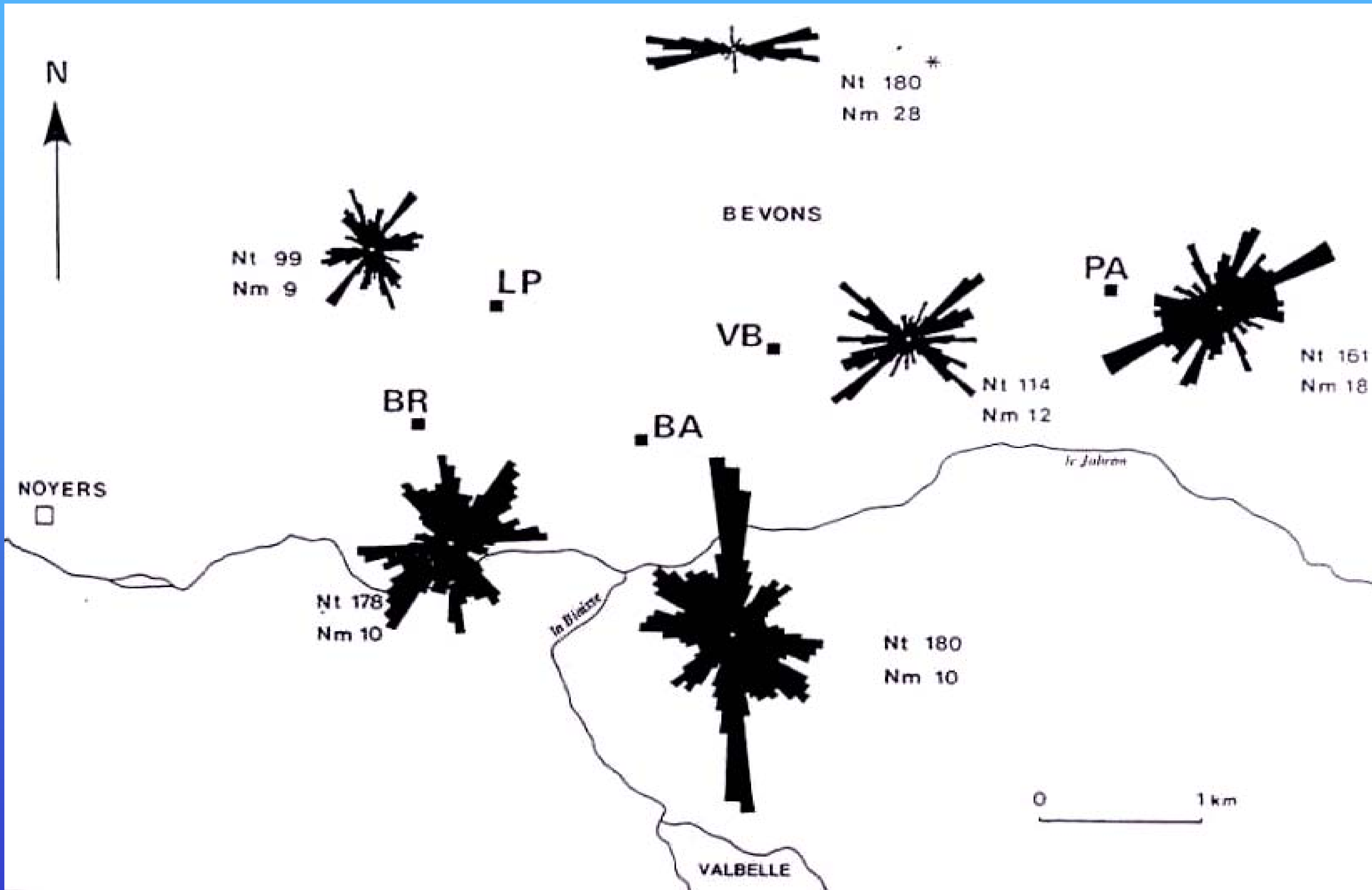
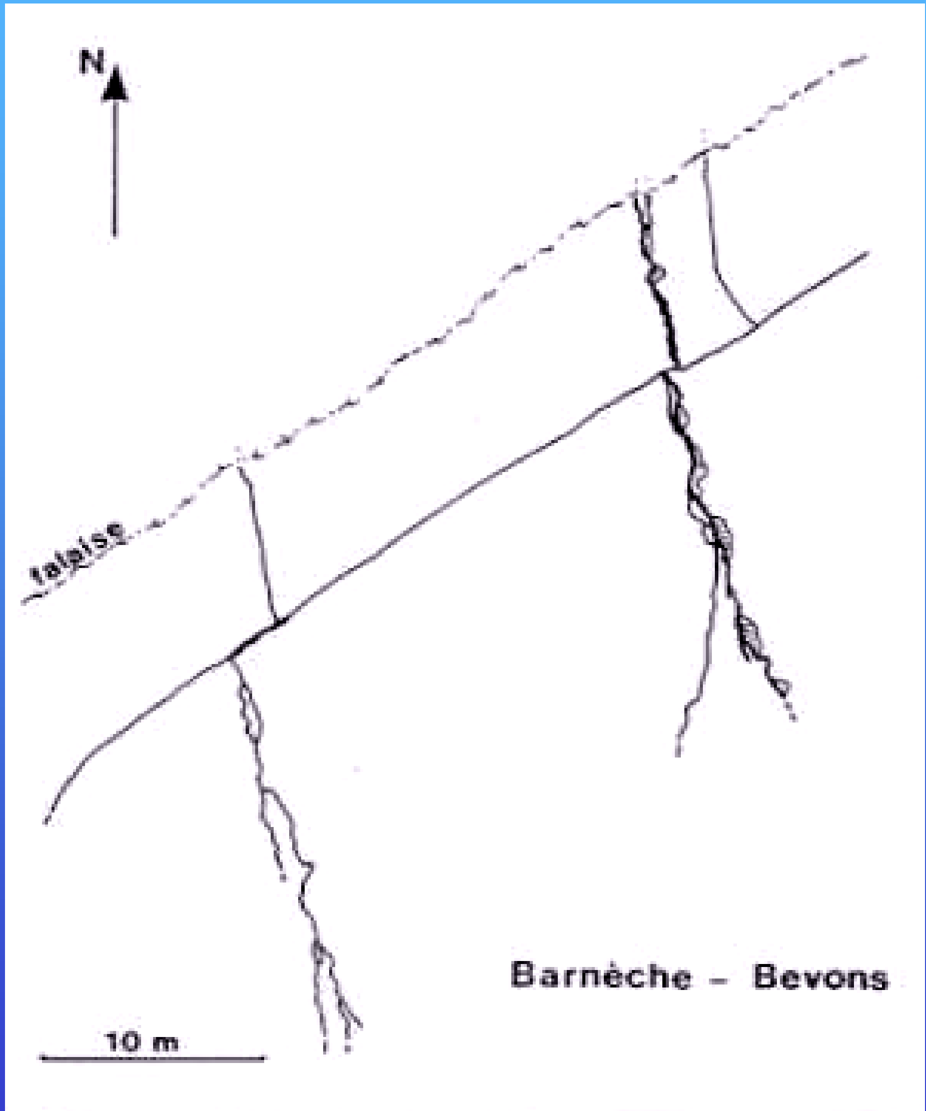
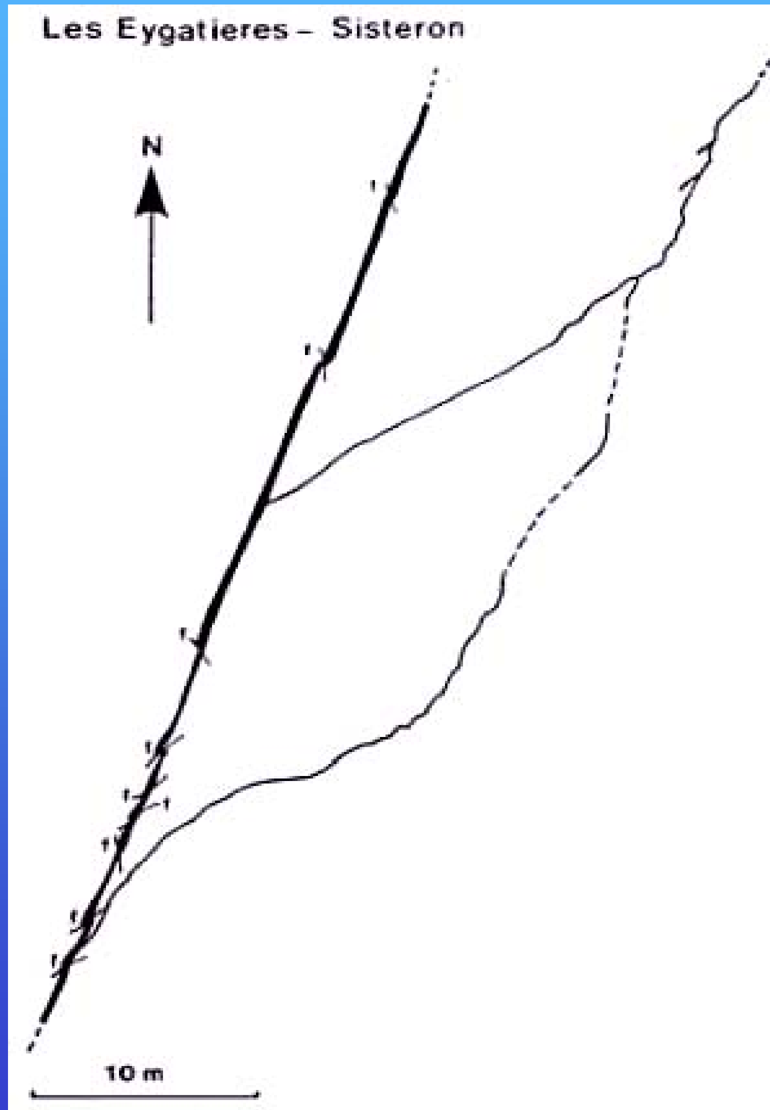
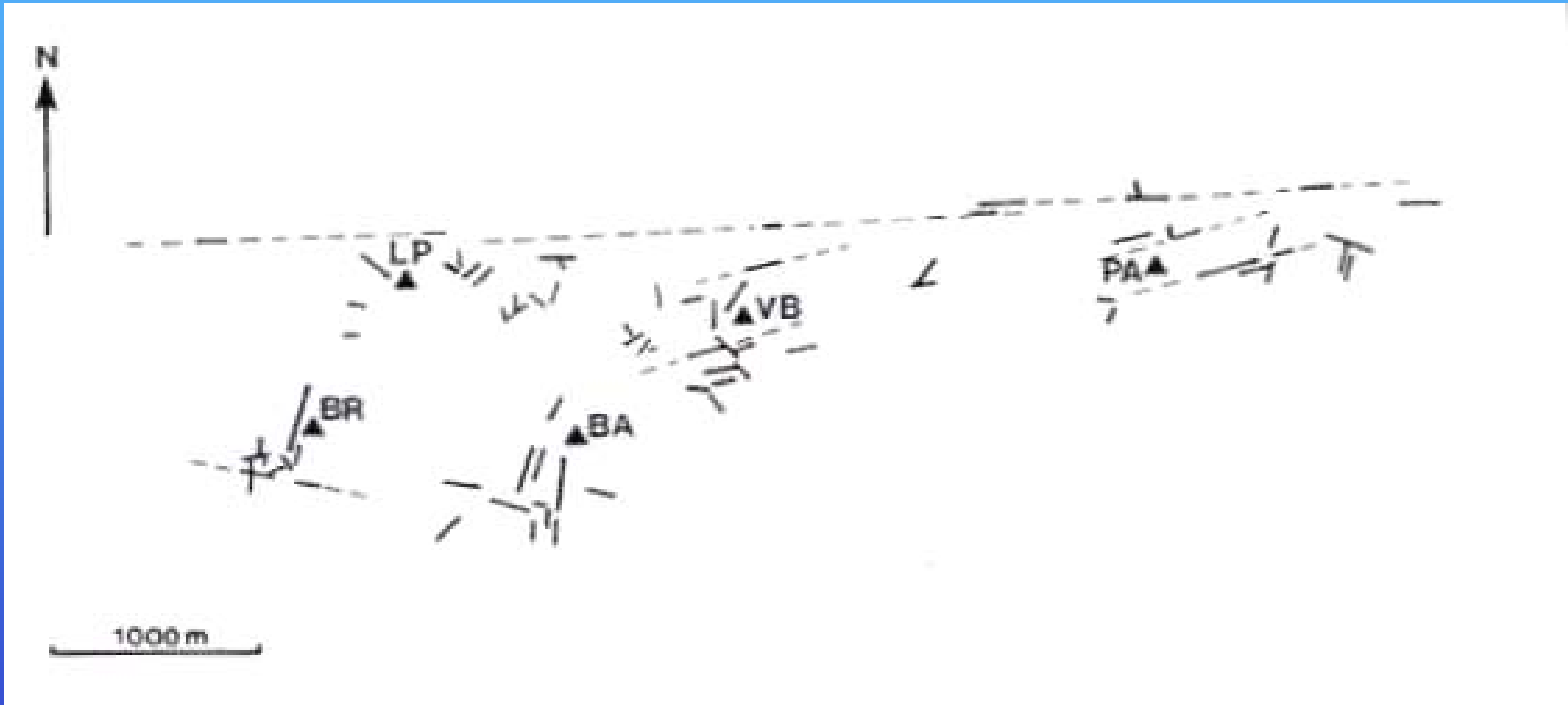


Dykes network examples

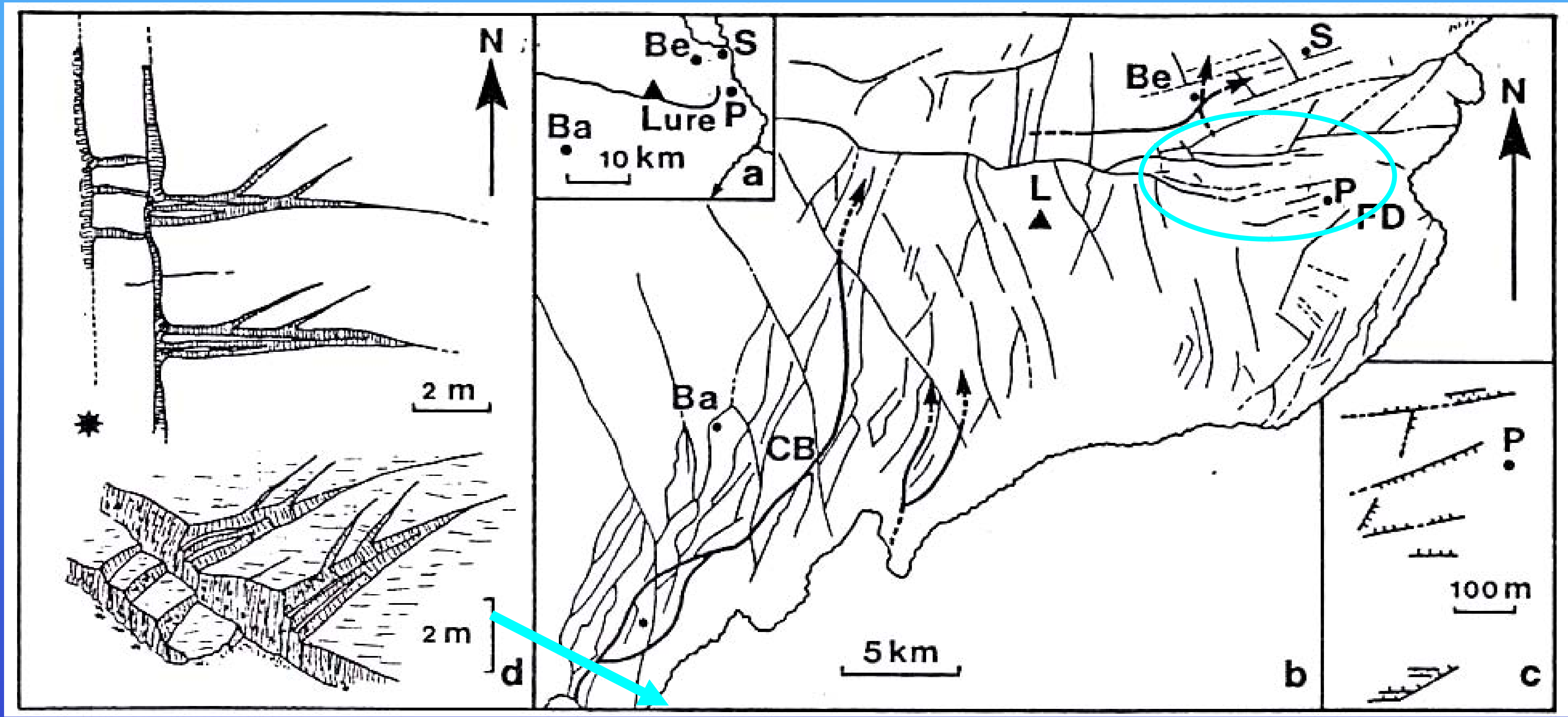




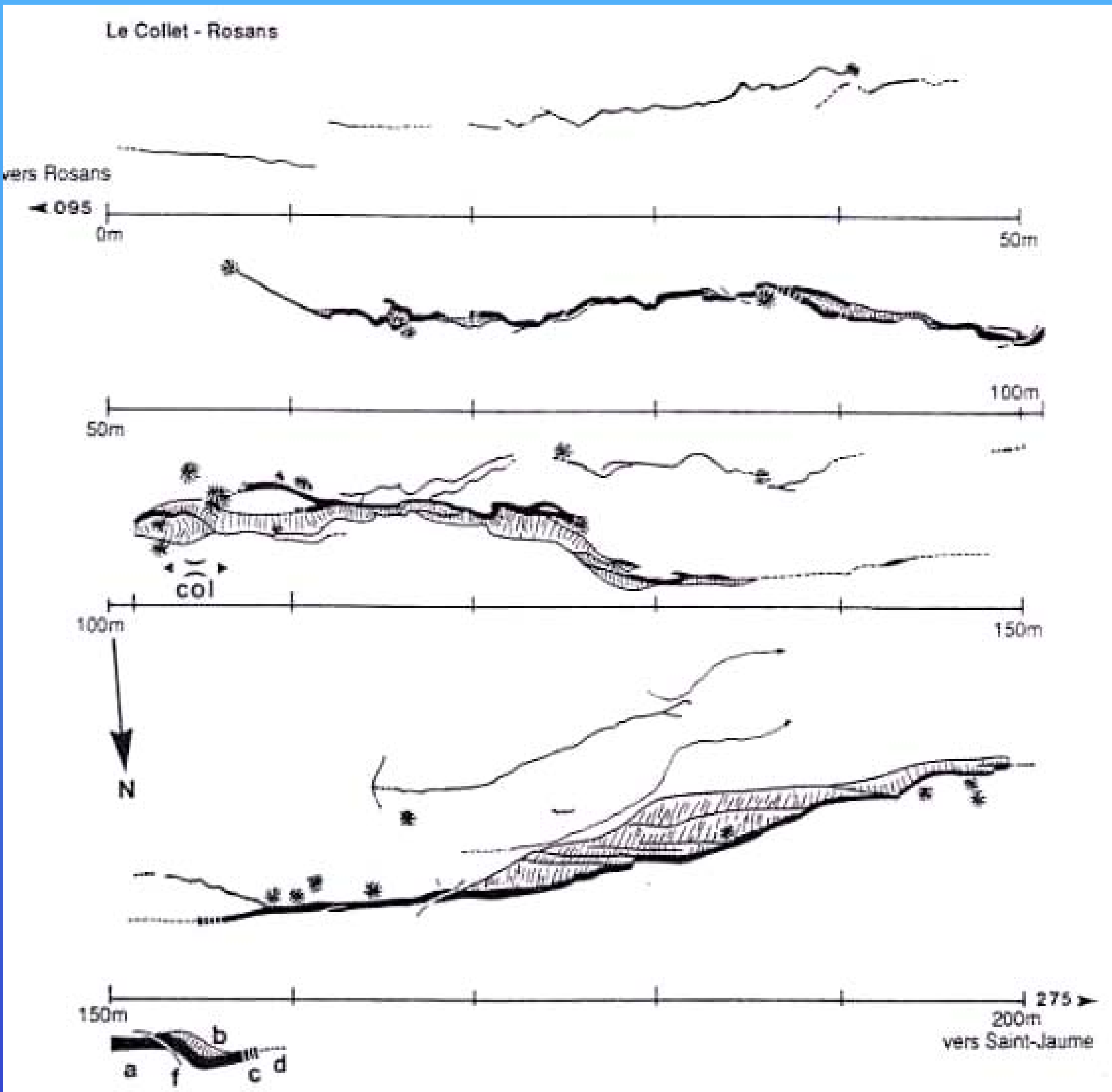
Mapping of the dykes network in the Bevons area



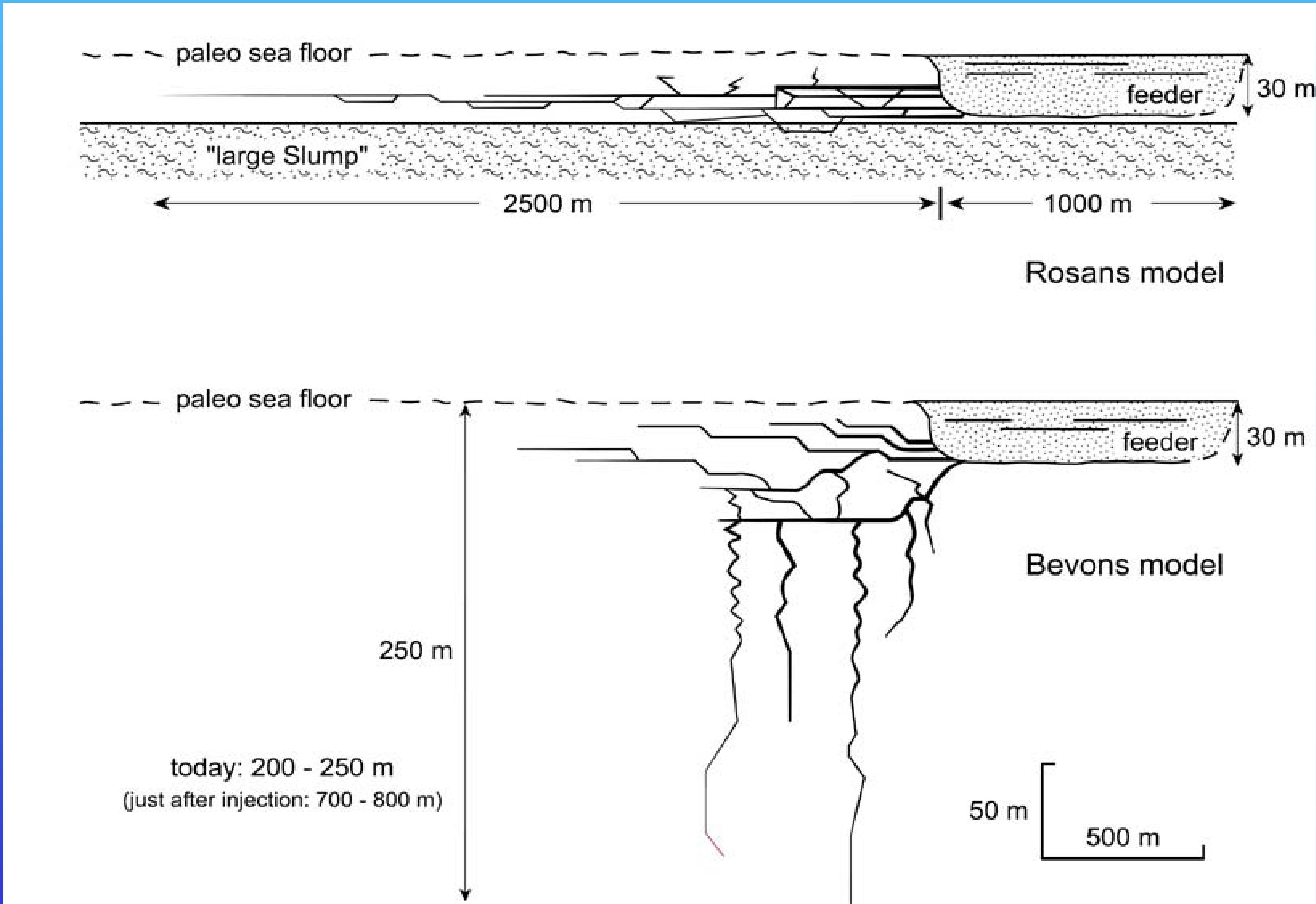
Synsedimentary tectonic setting



An example of dyke in Rosans area



Vocontian injections - feeder networks



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CONCLUSIONS

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VOCONTIEN
CLASTIC INJECTIONS

DISTAL PART OF
MASSIVE EROSIONAL SYSTEMS

PER DESCENSUM INJECTIONS
(under the palaeo-sea floor)

DIFFERENTIAL COMPACTION

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VOCONTIAN
CLASTIC INJECTIONS

FROM THE FEEDER

1- SILLS IN THE BANKS

2- DYKES FROM THE SILLS

NO DYKE (OR SMALL DYKES) UNDER THE CHANNEL INFILLS

A syndepositional model

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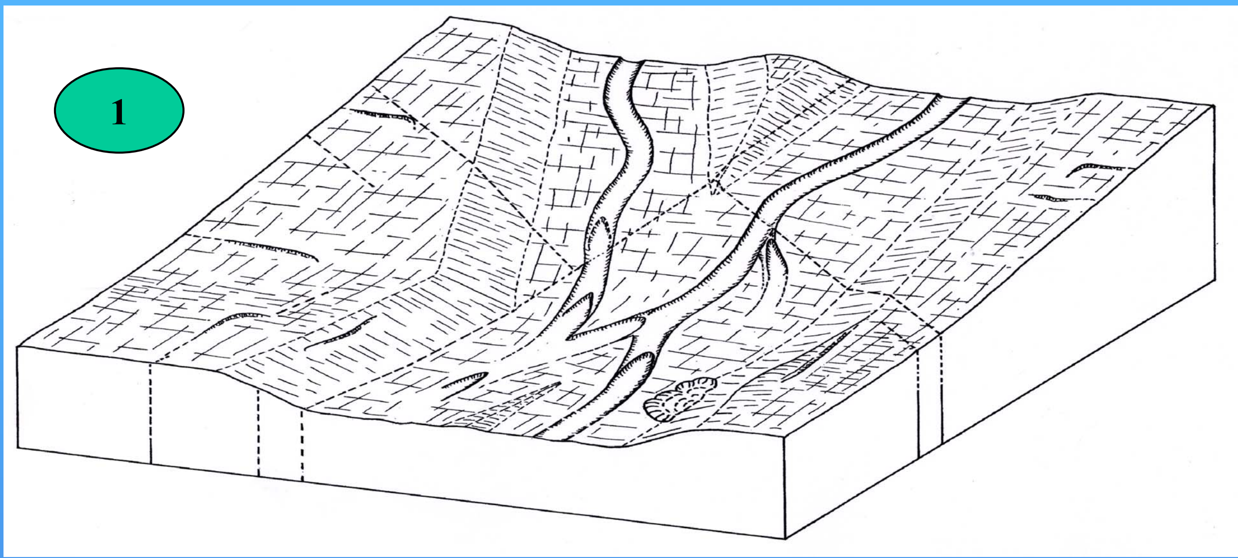

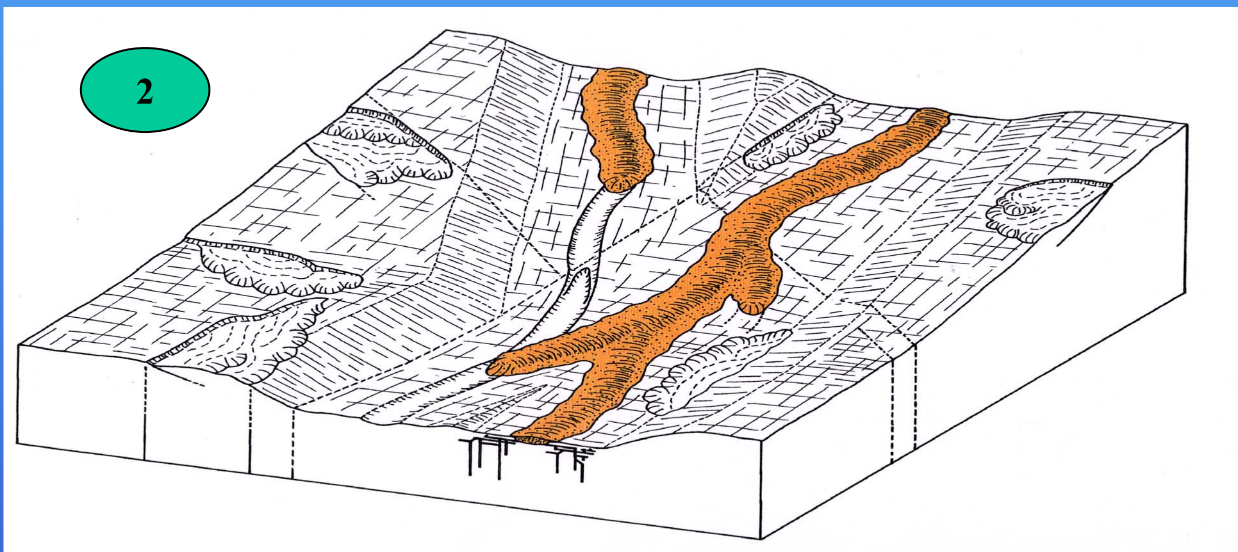

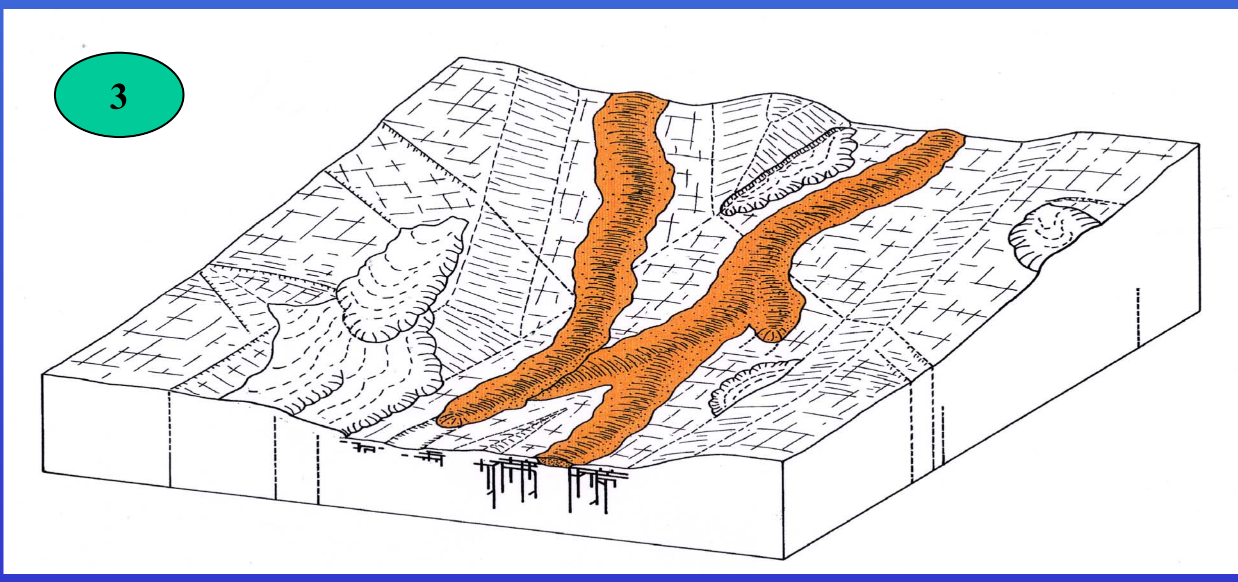

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VOCONTIAN
CLASTIC DYKES

EARLY FRACTURATION

1- DIFFERENTIAL COMPACTION

2- REGIONAL MORPHOLOGY

3- REGIONAL DEFORMATION

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VOCONTIAN
CLASTIC DYKES

SAND INJECTION

- **CHANNEL GEOMETRY**
  (Erosional features)

- **TURBIDITIC DEPOSITION**
  (Channel infilling)

Notes

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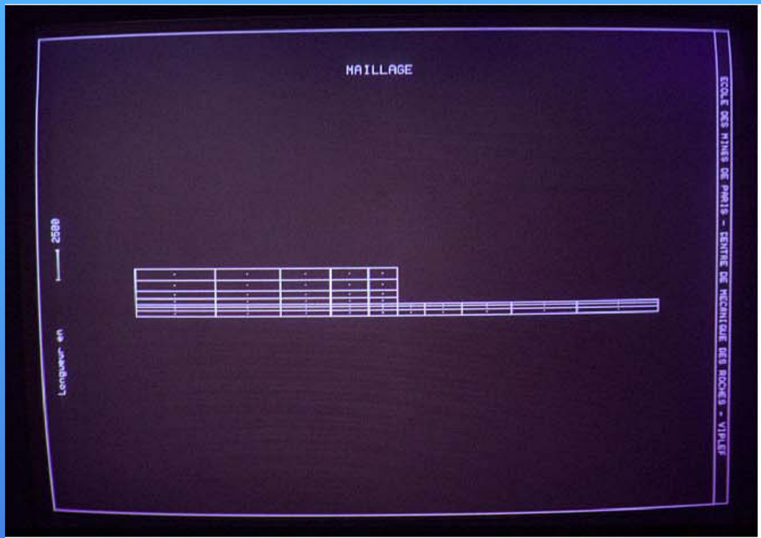
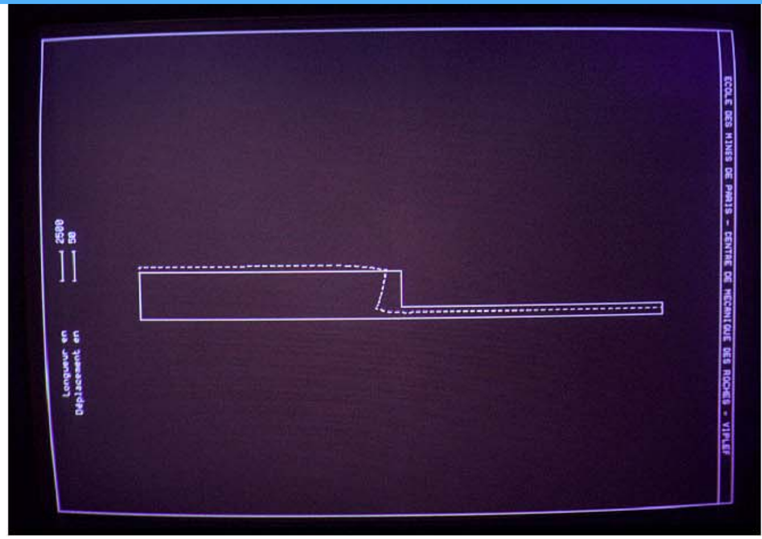
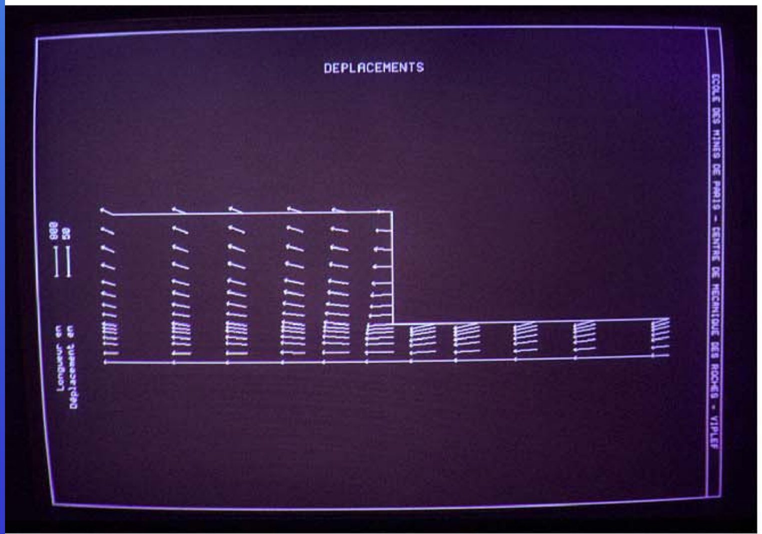
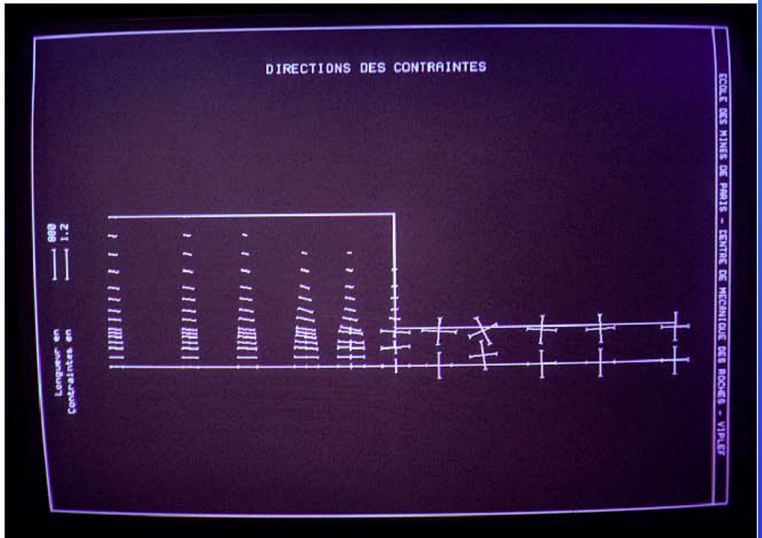
Mechanic simulations (1)









Mechanic simulations (2)

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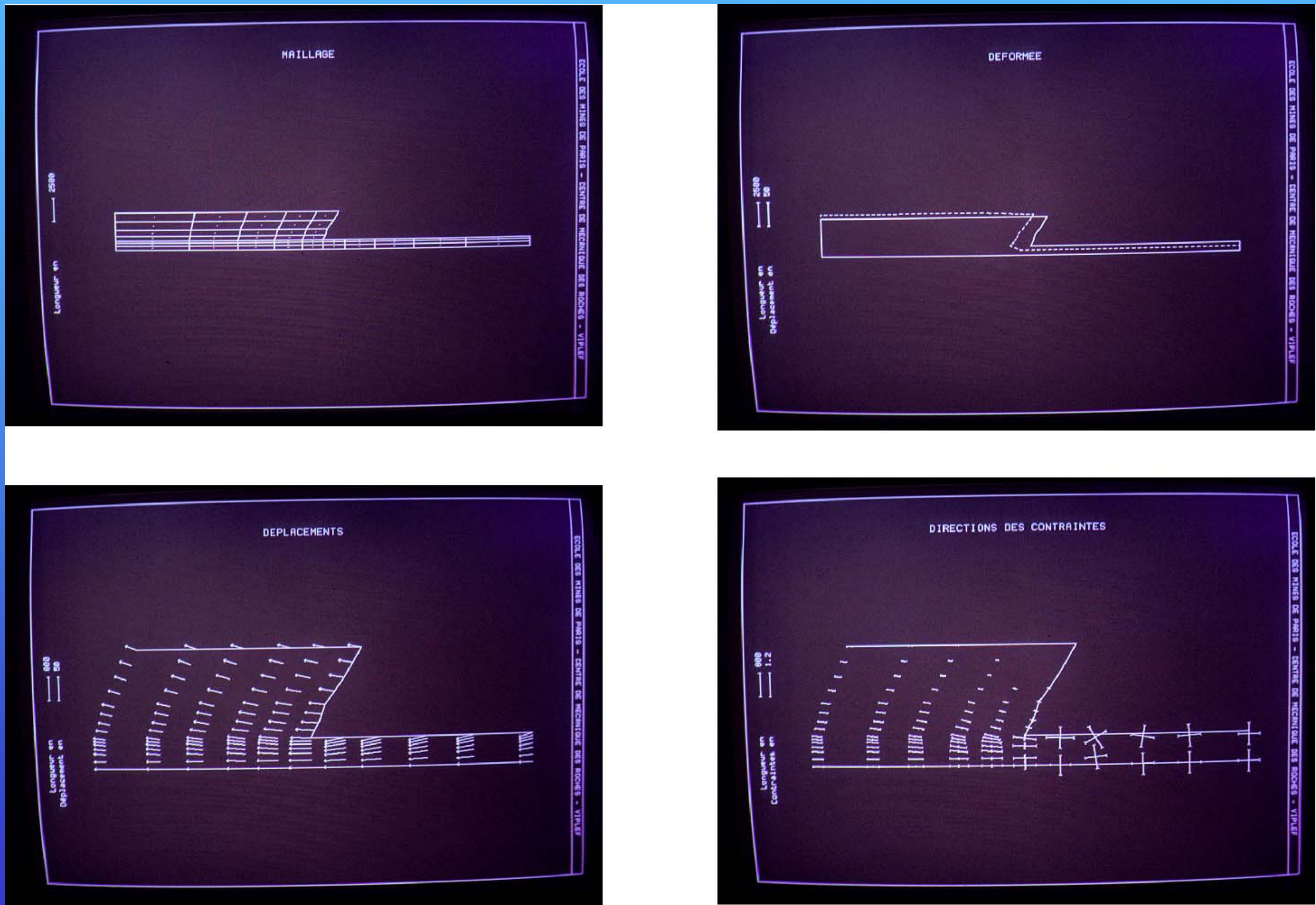
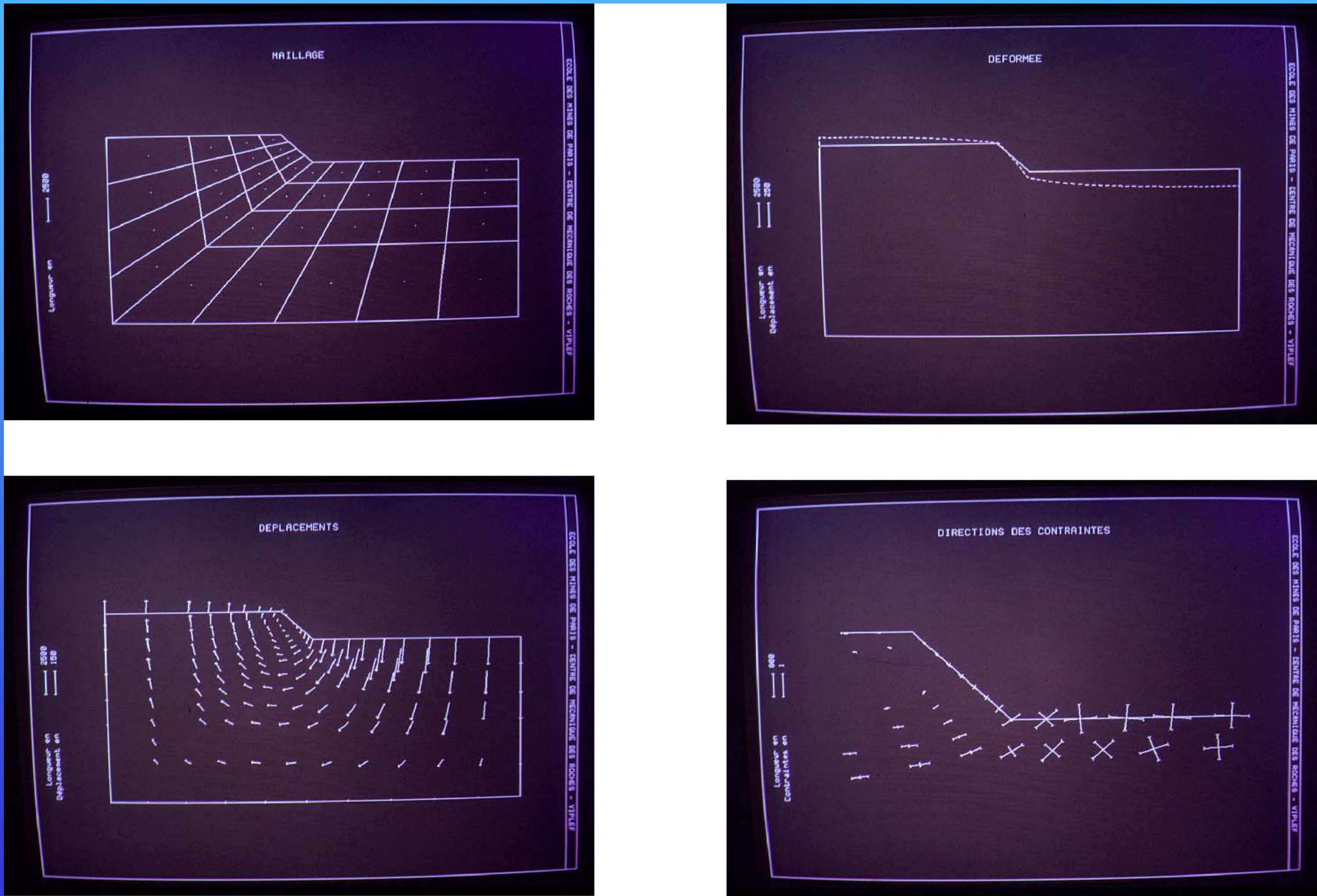


Mechanic simulations (3)



Mechanic simulations (4)

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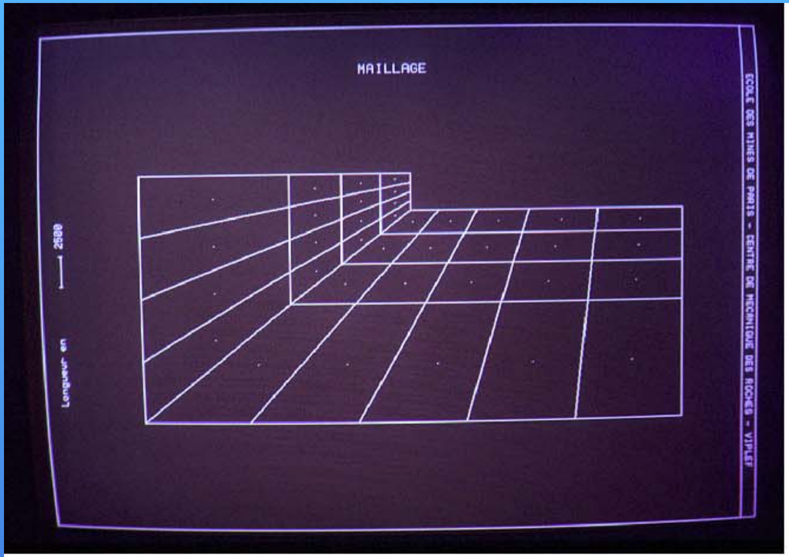

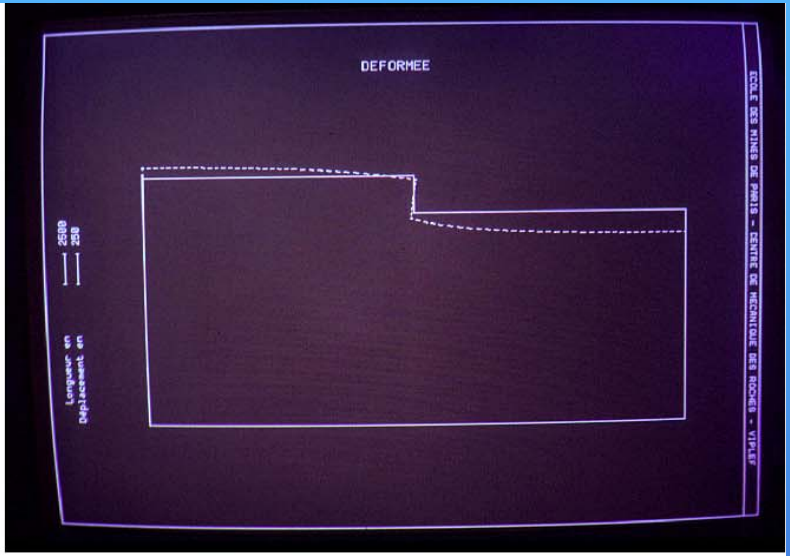

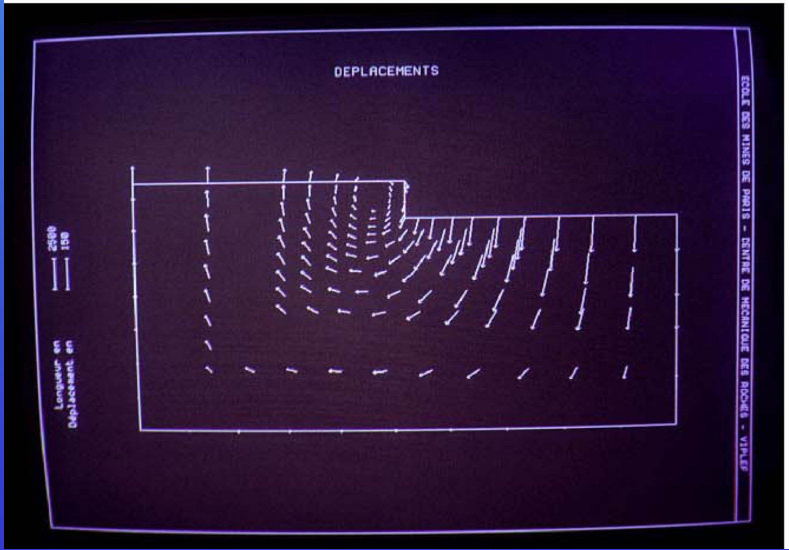

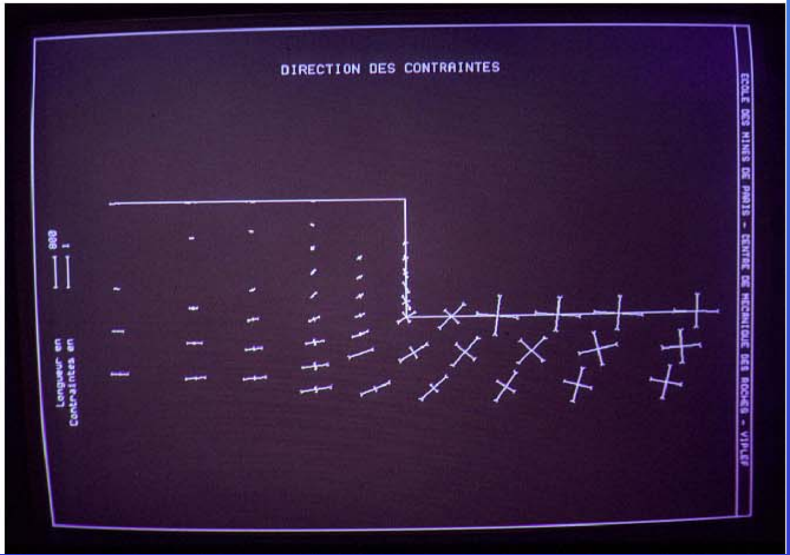

Vocontian injectites
a cartoon

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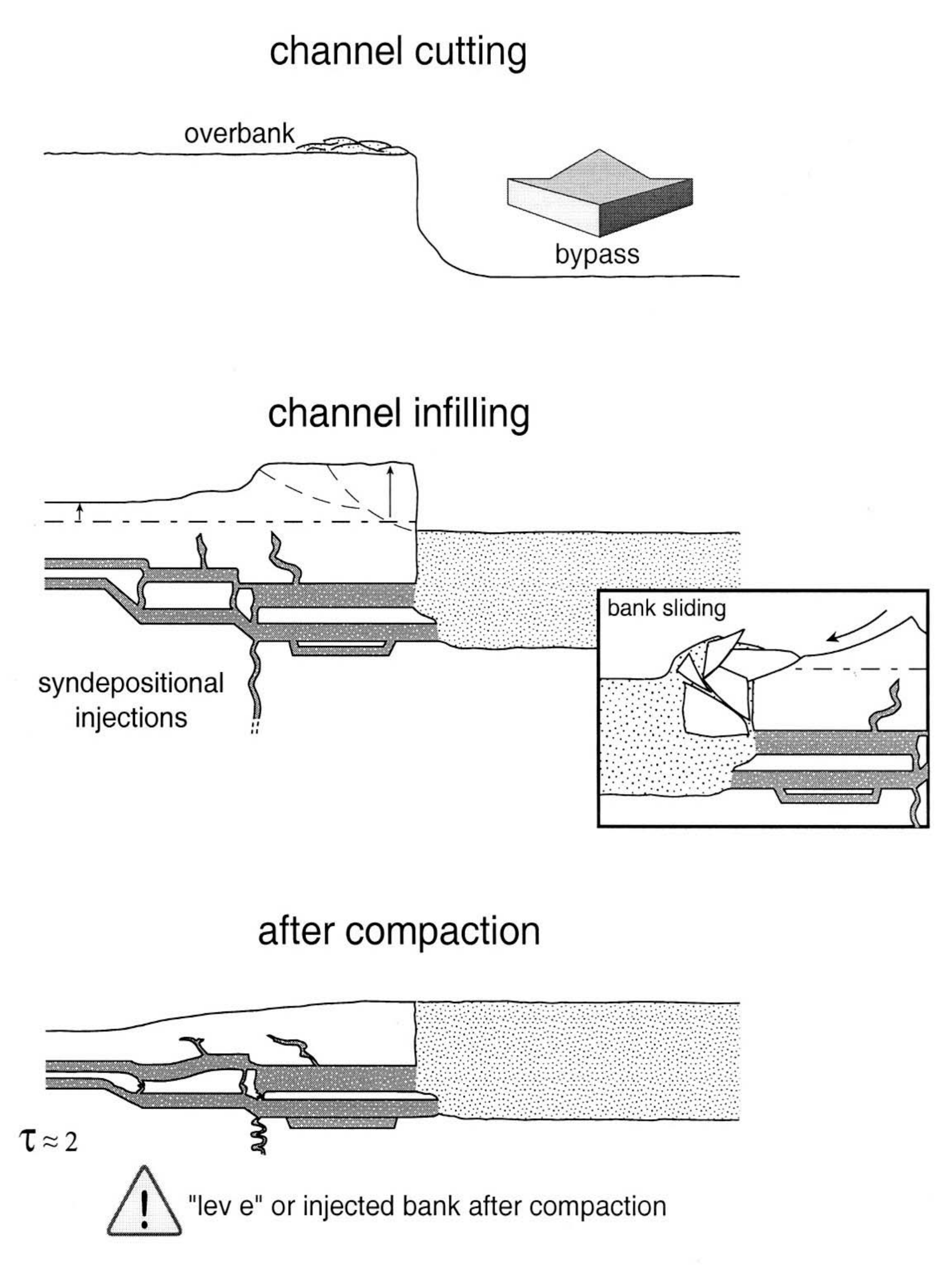

THE VOCONTIAN INJECTITE EMPLACEMENT:
A CARTOON

The clastic injection is an instantaneous event related to the “en masse” deposition of a massive turbidic event in a erosive channel: three successive stages can be easily define as (i) the shaly sediment mass immediately before the turbidite feeder coming, (ii) this mass during the “en masse” deposition of the turbidite feeder and (iii) the injectite network and the host formation after compaction, looking like the today exposures.

Stage 1: The cutting

The submarine floor is cut by a eroded sediment flow, for example a hyperpycnal flow similar to Nice Baie des Anges (Parize *et al*., 1989; Mulder *et al.*, 1996) even if the initial triggering can be inferred to sand liquefaction on shelf during storms or tsunami (Imbert *et al*., 1995). This flow is associated only to thin overbank deposits and the important cutting of the superficial sediments.

Stage 2: The infilling

The sudden deposition of a sandy turbiditic flow provokes the fracturing of the bank and the sand injection into them. It is mechanical possible as proposed by Eckert *et al.* (Ghent Conference, 2001; c.f. Parize, 1988; Parize *et al*., 2001, 2002). This deposition is inferred as “en masse” characteric of high density turbiditic flow *sensu* Lowe (1982) or granular flow *sensu* Mutti *et al*. (1999). This turbiditic event is associated with the bank undercutting and undermining. No argument allows to indicate that this event is completely separated from the previous one.

The injection into the bank provokes the uplift of them, called lift effect. An equilibrium state appears between the sand injected and the sand infilling the channel: this infilling cannot overtake the banks and the channel infilling and the injection stop certainly at the same moment. This uplift can induced a lateral warping from the edge of the bank. It can be associated to sliding or slumping similar to bank caving and brecchia facies can occure locally.

Stage 3: The today shape, after compaction

After overlying and compaction, the resulting shape is similar with today exposures. This shape looks like channel-levee complex. From seismic data, this form convergence can induice a misinterpretation: for example, the surface covered by the sand complex and the volume of sand reservoir are greater than previous.

VOCONTIAN CLASTIC INJECTITES

In their distal extremity, the Aptian-Albian deep water channelized massive sands of the Vocontian basin (S. E. France) are often associated with sand injections (Beaudoin & Friès, 1982, 1984; Beaudoin *et al.*, 1983; Parize, 1988; Friès & Parize, submitted).

Located in the S. E. France, the Vocontian basin belong to the external part of the Western Alps. It is related to a preserved area of the upper-slope domain (Friès & Parize, in press) of the western margin of the Cretaceous Tethys – Valaisan palaeo-Ocean. In this domain, during the Aptian and Albian stages, numerous and various gravity-driven deposits are interbedded in the pelagic marls (Friès, 1987; Rubino, 1989). They consist of

(i) thick, up to 100 metres, marly to sandy slumps and debrites,
(ii) centimetre to metre thick Bouma turbiditic sequences and
(iii) metre to several decametre-thick massive sands (Rubino, 1989; Friès, 1987; Imbert *et al.,* 1995; Friès & Parize, submitted).

The morphological reconstitution and the sedimentological analysis demonstrate that these deposits were controlled by synsedimentary faulting and halokinesis (Friès, 1987; Joseph *et al.*, 1989; Parize, 1988; Friès and Parize, submitted) within a confined setting: the turbiditic systems, 40 to 80 kilometres-long, are channelled from the shelf-edge to their distal part and the main accumulations are located in pounded basins (Friès, 1987; Parize, 1988; Rubino, 1989; Friès & Parize, submitted) as described on modern ocean margins.

The Bevons (Beaudoin & Friès, 1982, 1984) and Rosans (Beaudoin *et al.,* 1983) areas, in the Vocontian domain, present probably the most spectacular outcrops showing complex networks of clastic sills and dykes injected in a marly/limy host formation. Most injectites are found in the channel banks, lateral to the feeding channel infilling (Beaudoin *et al.,* 1985; Parize, 1988). Clastic injectites consist predominantly of sills and subordinate dykes. The

Vocontian injections

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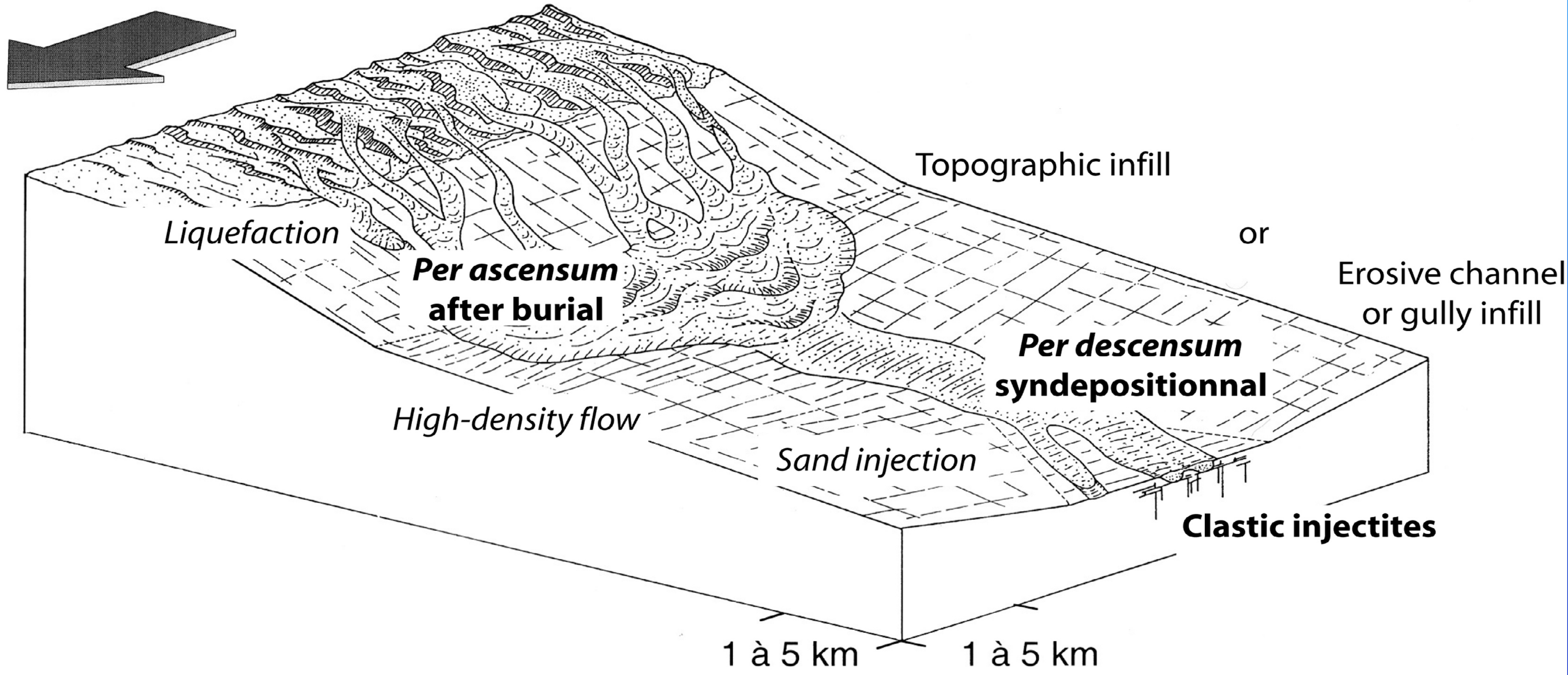

From Parize *et al*. and imbert *et al*. (1995)

sills are up to 10 metres thick in the vicinity of the connexion with the channel feeder; they thin out and die into marls 2 or 3 kilometres away from it (Beaudoin & Friès, 1984; Parize, 1988). Most dykes are injected from the sills rather from the channel itself: few small dykes can be located under the channel fill. They are more abundant within a few hundred meters away from the channel. Downward injections can reach 300 metres, whereas upward injections never reach the contemporaneous palaeo-sea floor (Beaudoin & Friès, 1982; Parize, 1988). Ptygmatic folding of the dykes by mechanical compaction indicates the amount of local post-injection compaction rate of shale which clearly shows that sand injection occurred prior to the burial (Schneider & Parize, 1989). Outcrops mapping shows that channel bank fracturing is contemporaneous to channel infilling (Beaudoin *et al*., 1985; Parize, 1988). This is an evidence of an early syn-depositional injection of the sandy material. Injections usually occur where the flow is slowed down by changes in topography ; they are commonly associated with large erosive scours.

Mechanical simulations of the compaction of the shaly and marly deposits on a structured substratum demonstrate that potential fracture zones can be induce by pre-existing decametre-high morphologies, such as fault escarpments, 400 metres below the palaeo-sea floor. Other mechanical simulations demonstrate that the infilling of an erosive channel (500 to 1000 metres wide and 20 to 50 metres deep) by one sandy event can explain the development of horizontal to oblique fractures in the banks of this channel. Therefore, the development of the sills is related to hydraulic fracturing by the water-sediment mixture (Beaudoin *et al.,* 1985; Parize, 1988).

The Vocontian clastic injectites model is related to sandstone sills and dykes injected during the infilling of erosional channels by high density *sensu* Lowe (1982) or granular *sensu* Mutti *et al.* (1999) turbiditic flows (Parize, 1988; Parize & Beaudoin, 1988; Imbert *et al.,* 1995; Parize *et al.,* 1999). This interpretation of sandy injections as dynamically related during the deposition of their feeder sandbody provides an alternative to the overloading – post-depositional - mechanism commonly proposed. Finally well-developed clastic sills might be interpreted to be the geometrical and physical equivalent of depositional lobes in confined settings.

PERPECTIVES

VOCONTIAN CLASTIC INJECTION EXAMPLE PROVIDE CERTAINLY GOOD GEOMETRIC ANALOGUES TO DEEP OFF SHORE CLASTIC INJECTED NETWORKS AND ELEMENTS TO DISCUSS GENETIC PROCESSES.

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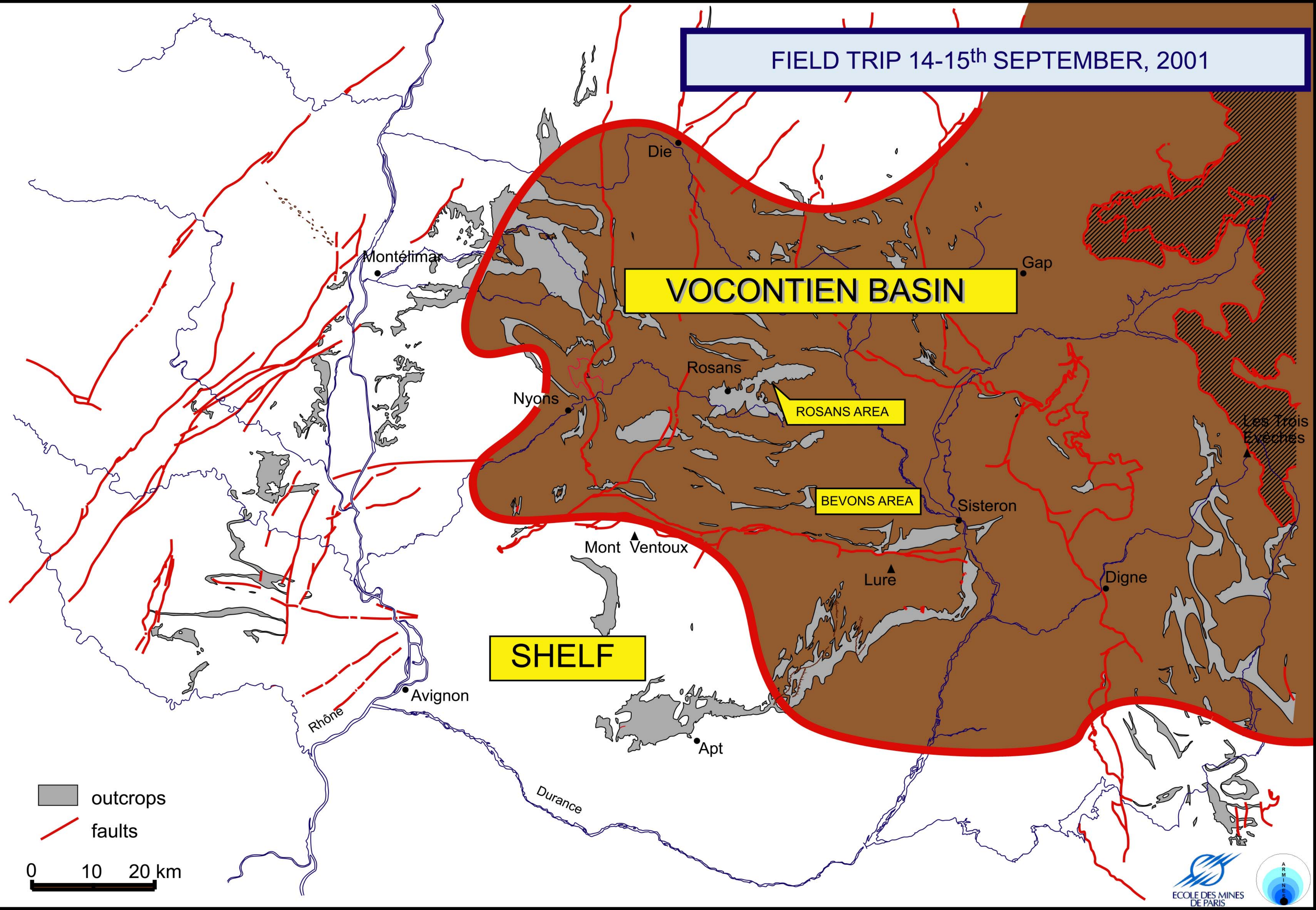

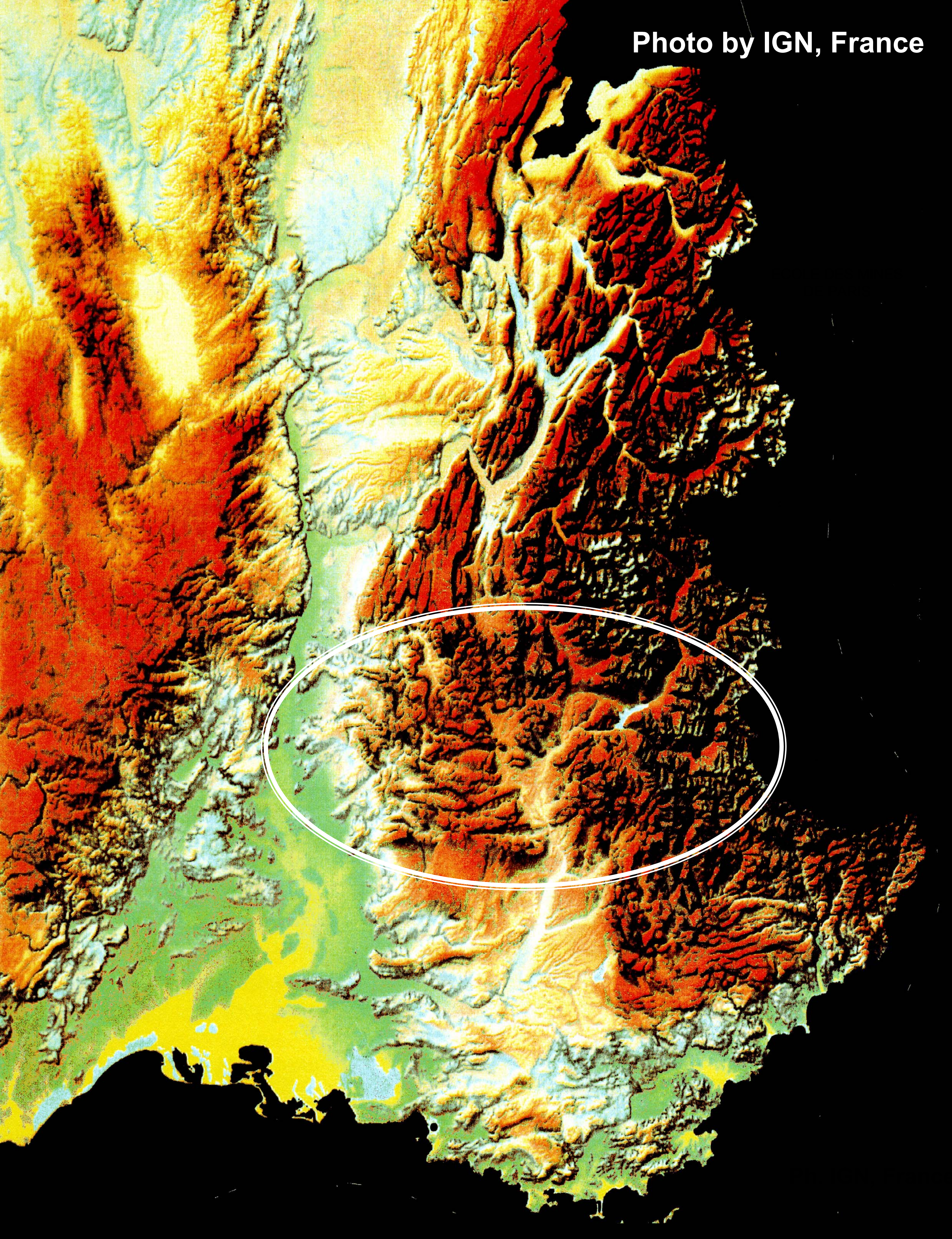

General location of Vocontian basin

Guide book field trip #2

Ghent International Conference

SAND INJECTIONS IN DEEP MARINE SHALE FORMATIONS IN SE-FRANCE

The Bevons and Rosans outcrops of Vocontian Aptian-Albian shales to marls ("Blue Marls" Formation) in SE France is one of the best exposed examples where sand injections in marine slope sediments can be observed and studied.

The injection network is clearly associated to massive channelised turbidites. Sand sills extend up to 3km from their source and are up to 12m thick; sand dykes can be up to 250m high and 5m wide. They display typical geometries as offshoots, lateral variations in thickness, and stepwise offsets. The sand injections are interpreted to be an indication for early fracturisation of soft deep-water mud and the orientation of the intrusions bears witness of the palaeo stress field.

Key words:
Dyke, dike, sill, clastic dyke, sandstone dyke, sandy injection, turbidite, compaction, outcrop, Alps, France, Cretaceous, passive margin, slump, debris flow, massive sand,