Mesozoic subsidence history of the European marginal shelves of the alpine Tethys (Helvetic realm, Swiss Plateau and Jura)

By Walter Wildi[1]), Hanspeter Funk[2]), Bernard Loup[1]), Edgardo Amato[1]) and Peter Huggenberger[1])

ABSTRACT

Based on a palinspastic restoration of the Helvetic realm, the Swiss Plateau and the Jura, the Mesozoic subsidence history of this area is reconstructed for about 50 stratigraphic sections and by taking into account the following parameters: age of the sediments, compaction corrected sediment thickness, depositional depth estimations and eustatic sealevel corrections for each lithological unit. The following major subsidence phases may be deduced from the geohistory diagrams:
— Triassic (mainly in the Jura),
— Early Jurassic (in the southernmost and western Helvetic realm only),
— early Middle Jurassic (in the Jura and the western part of the Helvetic realm),
— Early Cretaceous (in the southernmost and western part of the Helvetic and the Subalpine realms).

In the Jura, the Triassic and Middle Jurassic phases are probably due to intracontinental rifting following Late Variscan structures. The Early Jurassic phase due to extensional tectonics is locally well established in the Helvetic realm.

For the Late Jurassic and Early Cretaceous phases, the subsidence is mainly due to extensional tectonics on the northern margin of the Tethys.

ZUSAMMENFASSUNG

Die Mesozoische Subsidenzgeschichte des abgewickelten Helvetischen Raumes, des Schweizerischen Mittellandes und des abgewinkelten Jura wird anhand von 50 stratigraphischen Profilen rekonstruiert, wobei folgende Parameter berücksichtigt werden: Alter, für die Kompaktion korrigierte Sedimentmächtigkeiten, die Ablagerungstiefen sowie die eustatischen Meeresspiegelschwankungen. Folgende regional wichtige Phasen rascher Subsidenz können aus den Subsidenzdiagrammen abgeleitet werden:
— Trias (hauptsächlich im Jura),
— Lias (nur im südlichsten und im westlichen Helvetischen Raum),
— Früher Dogger (im Jura und im westlichen Teil des Helvetischen Raumes),
— Malm (in allen paläogeographischen Räumen),
— Frühe Kreide (im südlichsten Teil und im westlichen Helvetischen Raum, sowie im Subalpinen Raum).

Im Jura können die triadische und mitteljurassische Phase vermutlich durch die Remobilisierung spätvariszischer Strukturen erklärt werden. Die liasische Extensionsphase ist im Helvetikum lokal gut dokumentiert.

Die spätjurassische und die frühkretazische Subsidenzphase können durch Extensionstektonik am Nordrand der Tethys erklärt werden.

[1]) Département de Géologie et Paléontologie, 13, rue des Maraîchers, 1211 Genève 4, Switzerland (Swiss National Science Foundation project n° 2.107–0.86).

[2]) Geologisches Institut ETH-Zentrum, 8092 Zürich, Switzerland (Swiss National Science Foundation project n° 2.155–0.86).

RÉSUMÉ

L'histoire de la subsidence du Domaine helvétique, du Plateau Suisse et du Jura au cours du Mésozoïque est analysée sur une cinquantaine de coupes choisies comme type, pour lesquelles les paramètres suivants ont été pris en considération: âge des sédiments, épaisseurs décompactées, profondeur de dépôt et variations eustatiques. Plusieurs épisodes de subsidence régionale plus rapide se déduisent de notre étude. Ils peuvent être datés respectivement du
— Trias (essentiellement dans le Jura),
— du Lias (dans l'extrémité sud et la partie occidentale du Domaine helvétique),
— du début du Dogger (dans le Jura et la partie occidentale du Domaine helvétique),
— du Malm (dans tous les domaines),
— du Crétacé inférieur (dans l'extrémité méridionale et la partie occidentale des domaines helvétique et subalpin).
Dans le Jura, les deux phases du Trias et du Dogger s'expliquent par la remobilisation de structures tardivarisques. La phase liasique expliquée par la tectonique distensive, affecte bien des parties du Domaine helvétique.
Au Malm et au Crétacé inférieur, la subsidence de la région étudiée est essentiellement commandée par la tectonique distensive de la marge nord de la Téthys.

1. Introduction

The outcrops of sediments and basement rocks in the Alpine ranges offer particularly interesting conditions for the study of the subsidence of former sedimentary basins, despite the deformation of the lithosphere by the Alpine orogenic events. Recent studies on the Mesozoic margins of the alpine Tethys have stressed extensional tectonics as a major reason of basin subsidence:
— On the southern margins, i.e. in the South-alpine and Austroalpine, the Umbrian and Tuscan units, sedimentary facies types and changes in sediment thickness indicate east-west crustal extension from the Triassic to the Cretaceous (BERNOULLI et al. 1979a, b, CASTELLARIN 1980, WINTERER & BOSELLINI 1981, EBERLI 1988).
— On the European margin of the Western Alps, a tilted block model explained by extensional tectonics has been proposed by French geologists, following the observation of paleofaults, scarp breccias and reworked volcanics (synopsis in BOILLOT et al. 1984, LEMOINE & TRÜMPY 1987, GRAND 1988).
— On the European margin of the Central Alps, north-south extensional tectonics has been evoked since a long time to explain subsidence of the Helvetic shelf (bibliography in: LEMOINE & TRÜMPY 1987), and a detailed subsidence history analysis by FUNK (1985) has confirmed the stepwise and blockwise sinking of the Helvetic realm in Eastern Switzerland. This mechanism has been generalized to explain the development of the entire northern margin of the Alpine Tethys, including the north-Penninc through and the mid-Penninic platform (Briançonnais s.1.) (FUNK et al. 1987, Figs. 1 and 2), and oblique strike-slip movements may have contributed to this evolution (KELTS 1981, WEISSERT & BERNOULLI 1985, TRÜMPY 1988).

A quantitative estimation of the amount of Mesozoic crustal stretching of the Helvetic area provides low values of about 5 to 10% (WILDI et al. 1987).

Similar sediment facies types and sediment thicknesses as in the northern part of the Helvetic area are found below the Swiss Plateau and in the Jura mountain belt, up to more than 100 km to the north of the Helvetic realm, on the Variscan borderland of the European continent. However, in these sedimentary realms, other subsidence mechanisms than extension may also be expected.

This paper tries to contribute to a better understanding of the Mesozoic subsidence history of the European borderland of the Tethys to the north of the Central Alps. On the basis of a compilation of literature data, it describes the history of cumulative subsidence of the base of the Mesozoic sediments. The results are discussed and qualitative evolutions of possible subsidence mechanisms are proposed.

2. Palinspastic reconstruction of the sedimentary realm

The geological units considered here are (Fig. 1):

— The *stable European foreland* and its sedimentary cover: the Tabular Jura, the massifs of the Black Forest and the Vosges, the Rhine graben, the Bresse graben and the small block of Ile Crémieux, in the south-west of the Folded Jura;

— The *Folded Jura mountain belt,* where the Mesozoic and few Cenozoic sediments have been folded and sheared off from their substrate and thrusted to the north and west, at the end of the alpine orogeny (LAUBSCHER 1965);

— The *Swiss Plateau,* where the Mesozoic sediments and their thick cover of Oligo-Miocene Molasse have been dislocated to the north by "décollement" in the Triassic evaporites ("substitution de couverture"). They have only been slightly deformed during the folding of the Jura mountains;

— The *Subalpine Molasse:* it consists of thrust slices of Oligocene Molasse along the Alpine border. The northern part of the former Mesozoic substrate is known by the gas well of Entlebuch 1 (VOLLMAYR & WENDT 1987);

— The *External massifs* of the Aare and the Aiguilles Rouges and their sediment cover, both thrusted on the top of the European foreland (PFIFFNER 1986);

— The *Helvetic nappes,* including Permian (so called "Verrucano") to Early Oligocene ("Helvetic Flysch") sediments.

— The *Subalpine chains* are the parautochtonous folded sediments in the front of the External massifs to the south-west of the Arve valley; the ranges of the Bornes and the Aravis are examined here.

In the palinspastic restoration of the sedimentary realm during the Mesozoic (Fig. 2), the *European foreland* is considered as unchanged and stable with respect to its present day geometry. This is a simplified solution, since east-west extension occurred in the Rhine graben and the Bresse basin up from the Eocene/Oligocene boundary (BERGERAT 1985).

The *Folded Jura* mountain belt was restored using existing information about kinematics of thrusting as well as indications concerning the direction of the main thrusts represented in Fig. 2 by arrows.

For the eastern part, we adopted LAUBSCHER's (1965) rational model for the shortening of the sedimentary cover. According to this model, shortening is zero at the eastern end of the fold belt (Regensberg, Fig. 1) and it would be about 17 km in the transect of Biel (I on Fig. 2) (for slight modifications, see LAUBSCHER 1986).

In the central part of the Jura ranges, to the north of Lake Neuchâtel, no indication for large overthrusts can be deduced from surface geology. However, the considerable depth of the magnetic basement of 1500 m below sea-level in the Doubs valley and 2500 m below sea-level below the town of Neuchâtel (KLINGELE & MÜLLER 1987) would allow tectonic doubling of the Mesozoic sequences between the syncline of La
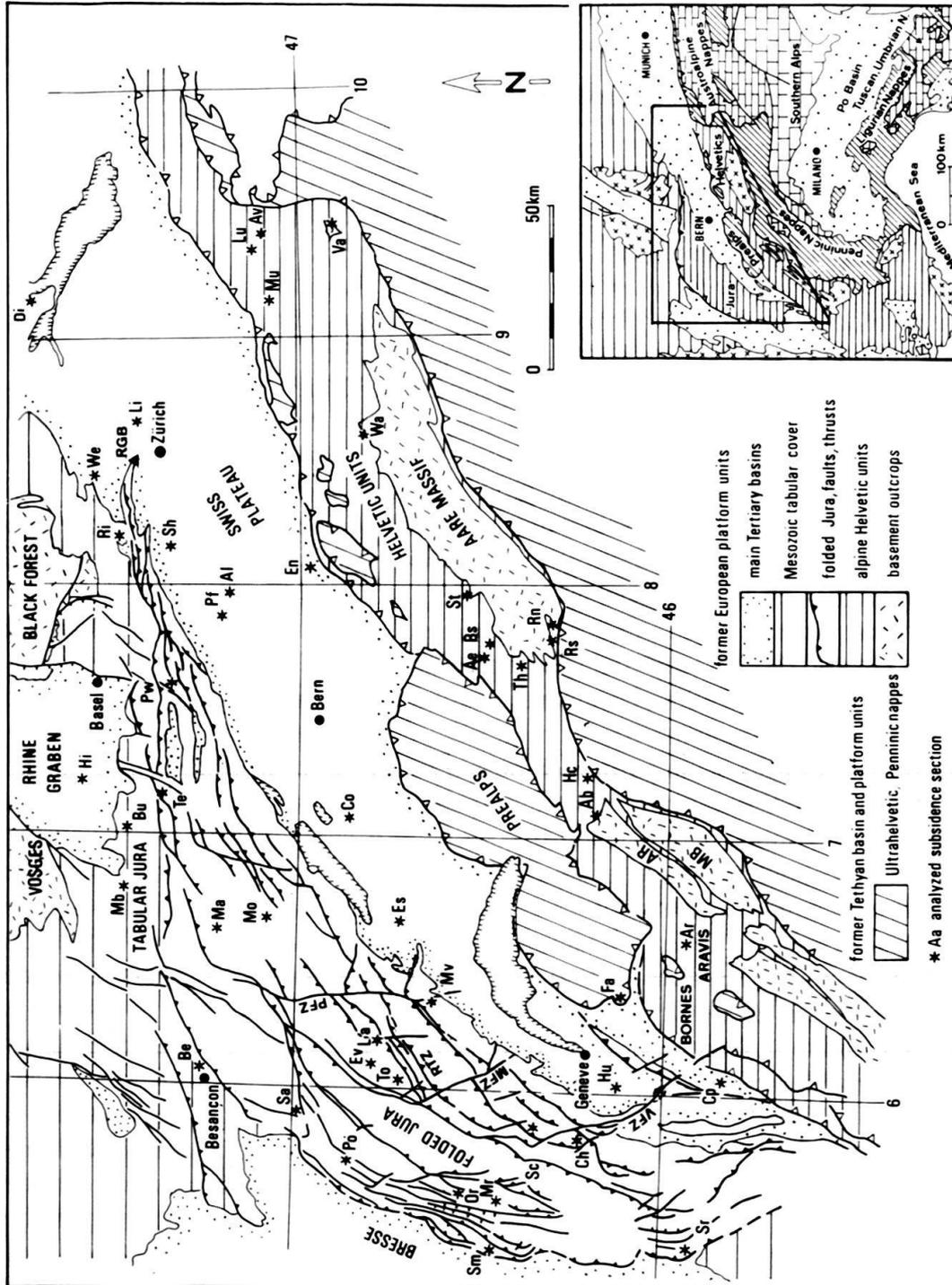
Fig. 1. Simplified map of the external part of the Alps, of the Swiss Plateau and the Jura; position of subsidence sections.
AR: Aiguilles-Rouges massif, MB: Mont-Blanc massif, MFZ: Morez fault zone, PFZ: Pontarlier fault zone, RGB: Regensberg (eastern termination of the folded Jura chain), RTZ: Risoux thrust zone, VFZ: Vuache fault zone.
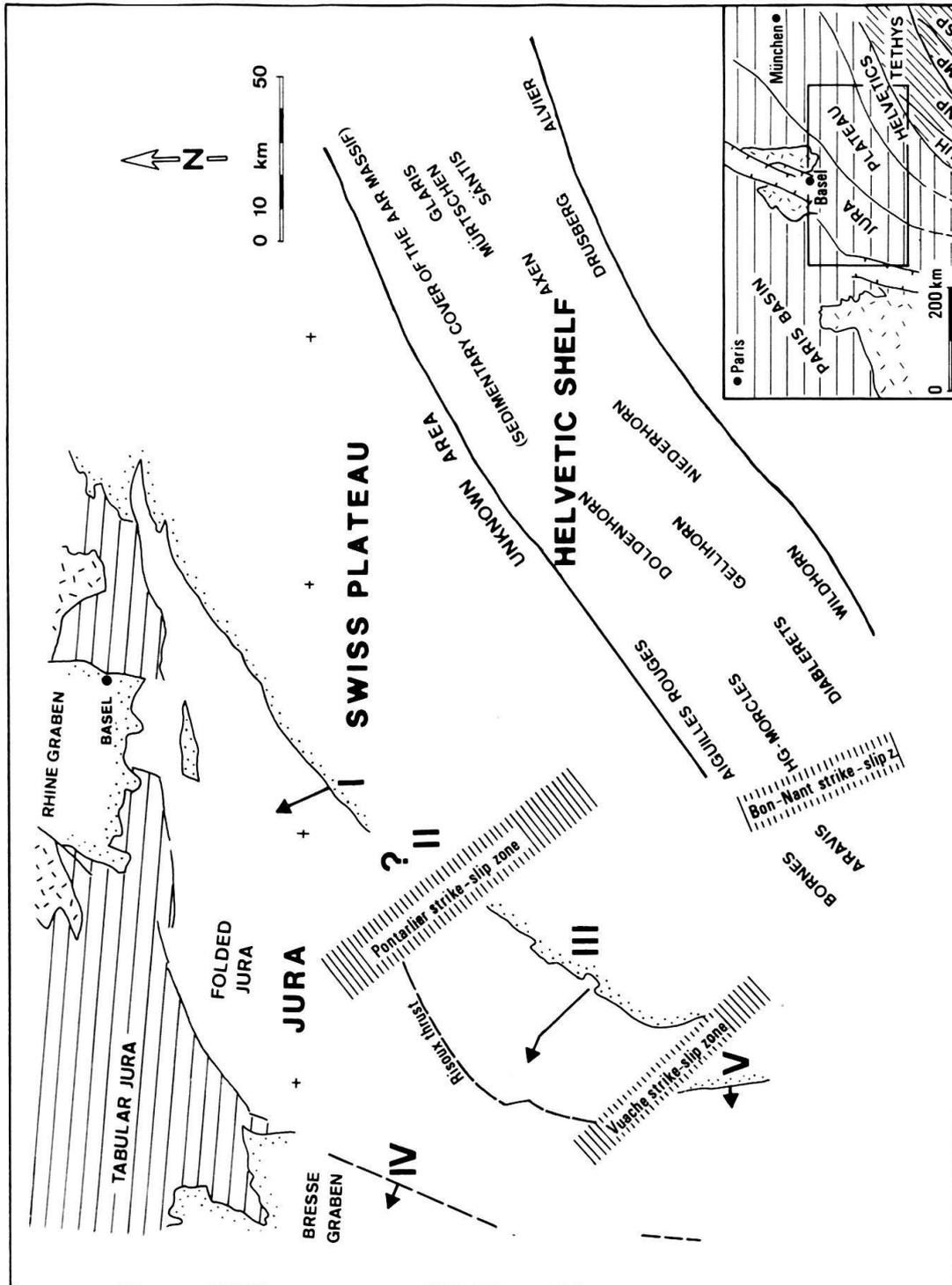
Fig. 2. Palinspastic map of the European marginal shelf of the alpine Tethys; Helvetic shelf, Swiss Plateau and Jura. HG: Haute Giffre, I–V: tectonic displacement in the central and western Jura (discussion in the text).

Chaux-de-Fonds and Lake of Neuchâtel, but this depends on the thickness of Upper Paleozoic sediments (at least 500 m after ZIEGLER 1982, plate 28). In Fig. 2 (II), a maximum shortening solution (about 18 km) is shown.

To the west of the Pontarlier strike-slip zone (PFZ on Fig. 1), overthrusting of more than 20 km has been indicated by the Risoux 1 well (WINNOCK 1961). Two north-looking ramps have been proposed by LAUBSCHER (1965) to explain this shortening of the sediments. On the other side, RIGASSI (1962) suggests one north-facing and one south-facing overthrust. One single SE-NW overthrust has been indicated by AUBERT (1971). This latter solution has been adopted in Fig. 2.

Small scaled ramps relayed by anticlines and synclines characterize the near surface geology in the high chain of the Jura between the Morez fault (MFZ on Fig. 1) and the Vuache strike-slip fault zone (VFZ on Fig. 1). The magnitude of overthrust in the frontal chain (Reculet-Cret d'Eau) along the Valserine valley develops from zero in the Col du Faucigny area to about 2 km near the western termination of this chain. The basement depth is at about 2500 m below sea-level below Lake Vouglans (lateral projection of the Valempoulière well, Mo-section in Figs. 4 and 5). In the hypothesis of a regularly rising Paleozoic/Mesozoic interface, a total shortening of about 22 km may be postulated for this transect of the high chain of the Jura (III on Fig. 2).

The amount of the overthrust of the Jura on the *Bresse graben* has been estimated to be more than 5 km in the Lons-le-Saunier section (e.g. RICOUR 1956) and of 8 km in the St-Amour transect (IV on Fig. 2). This is in agreement with the published geological maps, but in contradiction to MUGNIER & VIALON (1986) who obviously over-estimated the amount of extensional tectonics in the Lons-le-Saunier transect.

In the Southern *Jura mountains* between Geneva and Ile Crémieux (V on Fig. 2), the total shortening is estimated at about 8 km, including the frontal overthrust on the Bresse graben and shortening by the anticlinal faulting.

The *Swiss Plateau*, flat lying Molasse and its sedimentary and cristalline basement is considered in Figs. 1 and 2 as a rigid plate.

The Mesozoic basement of the *Subalpine Molasse* as well as the area situated below the overthrust of the *External massifs* on the top of the European foreland, correspond to the "Unknown area" of Fig. 2 (see also FUNK 1988 (Fig. 1) and FUNK & WILDI 1989).

Palinspastic reconstructions of the *Helvetic area* were proposed and discussed by TRÜMPY (1969) and FERRAZZINI & SCHULER (1979). The reconstruction used in this paper is a slight modification of the compilation by FUNK (in: TRÜMPY 1980). Each coherent nappe has been unfolded perpendicular to the main fold axis. Following PFIFFNER's (1981) results, the southern nappes have been repositioned in a southern direction instead of a SSE-direction used earlier. The result of this change: the southern Helvetic nappes (Wildhorn, Drusberg, Säntis-Churfirsten) have been located more toward WSW relative to the northern coherent units ("Autochtonous", Infrahel-vetic units, lower Helvetic nappes). The reconstructions are mainly based on formations of Early Cretaceous age.

For the construction of subsidence curves, it is important to study profiles which represent one paleogeographic locality prior to the tectonic deformation. In tectonically undeformed or only slightly deformed areas, this demand is easily fulfilled. In a nappe area, as for example in the Helvetic nappes, an original "cake" of formations may be

disrupted by shearing and strain due to thrusting. Thus it is not easy to situate the stratigraphic continuation of one lithologic profile in another nappe (FUNK 1985, Fig. 2).

The reconstruction of palinspastic profiles through the Helvetic nappes of Switzerland by TRÜMPY (1969), ZWAHLEN (1986) and BURKHARD (1988) gives an excellent basis for finding paleogeographic "pin points".

In the *Bornes massif of the Subalpine Chains,* between the Jura, the prolongation of the Swiss Plateau and the Helvetic realm, reconstruction is based on recent structural work on the basement/cover relationship (GOURLAY 1984).

A change in the trasport direction, recorded in the External massifs of France (GOURLAY 1984) and in the Morcles nappe (RAMSAY et al. 1985), from NNW and NW to WNW and W could also be established in the frontal and central parts of the Subalpine chains by tracing individual thrusts laterally. The shortening of the Bornes (including the Aravis chain) in the NW direction is estimated to be of the order of 11 km (minimum shortening for Aravis about 4 km, VILLARS 1987; due to lack of information, this distance could be greater) and a maximum of 8 km in the WNW direction (at Lake Annecy). The main structures may be followed laterally in the NE direction to the right side of the Arve valley (Rocher de Cluse). Further to the NE, a N-S running structure separates the Aiguilles Rouges massif from the Belledonne Externe massif (GOURLAY 1984), which may be responsible for the axial depression in the Giffre valley (Bon Nant strike-slip zone of Fig. 2). For the paleogeographic positions of the Subalpine and the Haut Giffre realms relative to the realm of the External massifs, we followed the interpretation of GOURLAY (1984). As a consequence, Morcles cannot be correlated laterally with the Aravis domain (see also the discussion by LATELTIN 1988, Figs. 34, 35, for the Early Oligocene paleogeography).

3. Calculation of subsidence history diagrams: techniques and parameters

The technique of the construction of subsidence history in general has been discussed by VAN HINTE (1978) and the cases in the context of the Alps by FUNK (1985) and RUDKIEWICZ (1988). The following geological parameters were taken into account for the construction of the subsidence history of the European marginal shelf presented here:
— sediments thickness (present day thickness, compaction, erosion and deformation)
— dating of sediments
— depositional depth
— eustatic sea-level variations.

Literature data and few original field data were computed to obtain the subsidence of the lowermost lithological unit of Mesozoic age.

This computer program with a user's guide may be obtained for IBM compatible PC's from W. Wildi (see adress on first page of this paper).

a) Sediment thickness

The *thickness* of relatively homogenous lithological units were generally taken from literature, such as descriptions of wells, explanation notes to geological maps and regional or thematic publications (see Table 1).

SECTION	COORDINATES	MAIN REFERENCES AND OBSERVATIONS
Ab Arbignon	CH 571.3/115.2	TRüMPY 1945, MASSON et al. 1980; tectonic deformation based on HUGGENBERGER 1985
Ae Aermighorn	CH 621/154.5 and 595/124	ZWAHLEN 1986; Trias and Lias of the Ayent Zone as the stratigraphic base of the Aermighorn section (see MOSER 1987, ZWAHLEN 1986); tectonic deformation not considered : only few data available
Al Altishofen	CH 640/228	KOPP 1952, FISCHER & LUTERBACHER 1963
Ar Aravis	FR 923/107	DéTRAZ et al. 1986
Av Alvier	CH 748/219	see FUNK 1985
Be Besançon	FR 875/255	geological map of France 1:50000, Besançon
Bs Bundstock	CH 623.5/152.5	ZWAHLEN 1986 (cum biblio); tectonic deformation not considered : only few data available
Bu Buix	CH 568/258	SCHMIDT & BRAUN 1924
Ch Champfromier	FR 868/138	JAQUET 1966
Co Courtion	CH 572/189	FISCHER & LUTERBACHER 1963
Cp Chapery 1	FR 887/100	Report ESSO-Rep (dip corrected)
Di Dingelsdorf	CH 728/288	LEMCKE & WAGNER 1961
En Entlebuch 1	CH 651/202	VOLLMAYR & WENDT 1987
Es Essertines 1	CH 539/173	BüCHI et al. 1965
Ev Essavilly 101	FR 885/205	geological map of France 1:50000, Champagnole
Fa Faucigny 1	FR 911/132	Report ESSO-Rep
Hc Haut de Cry	CH 582/116 to 582/124	BONNARD 1926, BADOUX 1972, MASSON et al. 1980; compacted thicknesses before tectonic deformation based on HUGGENBERGER 1985
Hi Hirtzbach	CH 583/271	VONDERSCHMITT 1942
Hu Humilly 2	FR 885/130	Report SNPA
La Laveron 1	FR 892/203	geological map of France 1:50000, Mouthe
Li Lindau	CH 692/255	BüCHI et al. 1965
Lu Lüsis	CH 745/234	see FUNK 1985
Ma Maiche	FR 925/260	geological map of France 1:50000, Maiche

Table 1 (part A): Location, main references and observations to the subsidence sections.

Erosion after deposition is generally badly documented. We systematically introduced erosion between the end of the Cretaceous and the Eocene in places, where the sediments of Late Cretaceous age are missing in the Alps and in the eastern part of the Plateau and the Jura.

In the Alps, *tectonic deformation* may strongly affect the estimation of thicknesses of the incompetent formations (shales and marls). A value of about 33% tectonic thinning was e.g. recorded by RAMSAY & HUBER (1983) for the Lower Liassic shale units of the normal limb of the Morcles nappe near the root zone. Only few data are available for the thickness change due to pressure solution documented by different kind of stylolites. Experimental work on folding of alternating competent and incompetent units (HUGGENBERGER 1985) however shows that straight limb parts of competent units will not be influenced much by folding at low to intermediate strains. These results were independently confirmed by undeformed fossils on competent fold limbs in the Morcles region.

Tectonic deformation was only introduced as a paramter in the alpine units (indicated on Table 1, parts A and B). But as a consequence of our lack of information on thickness change within incompetent units in the absence of strain markers and as a consequence of the problem of consistency between different authors, these estimates should be considered as approximate.

Compaction and cementation: the dominant sediment types are carbonates and marls; evaporites are known in the Triassic Muschelkalk and Keuper, sandstones in the Lower and Middle Jurassic and terrigenous shales occur in the Middle-Upper Liassic/ Lower Middle Jurassic (Obtusus, Posidonia and Opalinus shale type lithologies) and in the Lower Cretaceous ("Marnes d'Hauterive", Palfris marls, etc.).

SECTION	COORDINATES	MAIN REFERENCES AND OBSERVATIONS
Mb Montbéliard	FR 925/275	geological map of France 1:50000, Montbéliard
Mo Morteau	FR 930/240	geological map of France 1:50000, Morteau
Mr Moirans-en-Montagne	FR 850/150	geological map of France 1:50000, Moirans-en-Montagne
Mu Mürtschen	CH 730/215	see FUNK 1985
Mv Mont la Ville	CH 520/165	geological map of Switzerland 1:25000, Cossonay
Or Orgelet le Bourget	FR 850/177	Report SNPA (Dompierre sur Mont well) and
		geological map of France 1:50000, Orgelet le Bourget
Pf Pfaffnau 1	CH 632/231	BüCHI et al. 1965
Po Poligny	FR 850/215	geological map of France 1:50000, Poligny
Pw Passwang	CH 615/246	geological map of Switzerland 1:25000, Passwang
Ri Riniken	CH 656.6/261.9	MATTER et al. 1988b
Rn Raron Nord	CH 633.3/132.6	SCHENKER 1946, DOLIVO 1982 (cum biblio),BAER 1959
		tectonic deformation mainly from DOLIVO 1982
Rs Raron Sud	CH 630/130	see Raron Nord
Sa Salins les Bains	FR 879/230	geological map of France 1:50000, Salins les Bains
Sc Saint-Claude	FR 870/140	MUDRY & ROSENTHAL 1977
Sh Schafisheim	CH 653/246	MÜLLER et al. 1984
Sm Saint-Amour	FR 833/157	geological map of France 1:50000, Saint-Amour
Sr Saint Rambert	FR 850/115	geological map of France 1:50000, St Ramberg-en-Bugey
St Stechelberg	CH 636/155.2	KREBS 1925, COLLET & PARéJAS 1931, BRUDERER 1924,
	to 633.3/154.6	MASSON et al. 1980; tectonic deformation : estimates from literature data
Te Mont Terry	CH 580/249	LAUBSCHER 1948; MAGE, personal communication
Th Torrenthorn	CH 620/134	SCHLäPPI 1978, 1980; FURRER 1939, 1962; v. Tavel 1937, BAER 1959,
	to 612/139.3	MASSON et al. 1980; tectonic deformation mainly from SCHLäPPI 1980
To Toillon 1	FR.881/196	geological map of France 1:50000, Champagnole
Va Vättis	CH 752/198	see FUNK 1985
Wa Wanneli	CH 688/187	NUESCH 1981
We Weiach	CH 676/268	MATTER et al 1988a

Table 1 (part B): Location, main references and observations to the subsidence sections.

The examination of thin sections from surface outcrops in the Jura mountain belt indicates low porosities of generally less than 5%, for sparitic calcarenitic rocks as well as for micritic, marly and shaly sediments. For the Helvetic samples, the porosity is even smaller than for those from the Jura. However, these low porosities cannot be explained by mechanical compaction due to burial alone, and major cementation has to be invoked. Unfortunately we do not know the time of cementation, nor the origin of the pore fluids which induced cementation: "sea-water, meteoric fresh-water, water squeezed from compacting shales and other formations, water with diverse histories that moves upward and outward through the compacting sediments of the subsiding basin" (BATHURST 1983, p. 349).

In the study area, only few data on compaction and cementation are available. We have therefore chosen a rather pragmatic model of sediment compaction during burial:
— compaction was computed after the formula of SCLATER & CHRISTIE (1980):

$$\Phi = \Phi_0 \times e^{-cz},$$

where Φ = porosity, Φ_0 = porosity of the unburied sediments, c = empirical parameter depending on the lithology and z = thickness of overburden (burial).
The following values have been introduced for c:
- bioherms and evaporites: $c = 1 \times 10^{-4} m^{-1}$
- arenites and calcarenites: $c = 3 \times 10^{-4} m^{-1}$
- marls, shales and mudstones: $c = 5.1 \times 10^{-4} m^{-1}$
- terrigenous shales: $c = 5.4 \times 10^{-4} m^{-1}$
(SCLATER & CHRISTIE 1980, SAWYER et al. 1982, HEIDLAUF et al. 1986)

— for Φo, i.e. the initial porosity of the sediments, we introduced a "compactible initial pore volume", corresponding to the maximum amount of pore space which may have been eliminated due to the mechanical compaction by burial (see e.g. Tables of PERRIER & QUIBLIER 1974). For the oosparitic "Urgonian" limestone (Barremian to Early Aptian) of the western Jura, which was never buried deeper than 100 to 500 m, and which is composed of about 50 to 60% of grains and of 40 to 50% of sparitic cement, the supposed "compactible initial porosity" is 5 to 10%. Inversely, for the terrigenous Aalenian Opalinus shales, 25 to 35% of compactible pore volume have been introduced, depending on the burial depth of the formation in the considered section. Similar or intermediate values between these two extremes (oolithic limestones with early diagenetic cements and shales) have been chosen for other lithologies, according to microscopic observations or to empirical suppositions. As a consequence of this method, the following systematic errors may have been introduced into the calculation of total subsidence:

— in our program, sediments with early diagenetic cementation are compacted in the same way as loose grain-by-grain sediments. This overestimation of the original sediment thickness is most probably less than 10%;

— for micritic and shaly sediments, the initial decompacted depositional thickness may be underestimated to up to 20%. This depends on whether the carbonate ions of the cements came from dissolution of the lithological units themselves, or whether they have been added to the rocks from outside.

— step-wise compaction during burial history was carried out lithological unit by lithological unit. This induced somewhat angular shapes of the geohystory diagrams in the cases of units of considerable thickness.

b) Dating of sediments

The dating of many of the borehole and outcrop sections used in this compilation is not based on biostratigraphic or radiometric evidence, but on lithologic and facies correlation. This very often can lead to dating errors due to heterochronous facies zones (GYGI & PERSOZ 1986, FUNKT 1989).

As stated in a earlier paper (FUNK 1985), the maximum error for the dated layers is of about 3 my for the Jurassic-Cretaceous part, up to 5 my for Triassic times.

For the timing of the sedimentary history we apply as far as possible the time scale and the biozonations of HAQ et al. (1987). The Triassic lithological units of germano-type facies are considered as isochrone in the whole area: Buntsandstein 250–240 my, Melsersandstein 240–238 my, Muschelkalk 240–231 my (subdivided into equal parts for Lower, Middle and Upper Muschelkalk), Keuper 231–215 my (Schilfsandstein 225–223 my) and Rhetian 215 to 210 my.

For the Lower, Middle and part of the Upper Jurassic (Lower Oxfordian), the ammonite zones of northern Europe are also valid for the Jura basin. Within the Middle to Upper Oxfordian lithological succession, the following dating concept has been chosen:

— "Argovian" (mainly the Wildegg Formation of GYGI 1969, 1986; GYGI & PERSOZ 1986): 149 my (base of Birmenstorf) to 146 my (top of Knollen-Beds, "Calcaires lités" of ENAY et al. 1984, 1988);

— "Rauracian" coral limestones in the southwestern Jura (Morillon beds of ENAY op. cit.): 146 my to 144 my, and starting at 148 my in the northern Jura (op. cit. and CYGI & PERSOZ 1986).

For the Late Kimmeridgian and the Portlandian, no detailed biostratigraphic data are available for the shallow carbonate platforms of the western and southern Jura and the Swiss Plateau. An almost isochronal unit at the top of these carbonates is the "Purbeckian" facies (MOJON & STRASSER 1987, STRASSER 1988), of 133–131 my (base of Cretaceous). Between 144 my and 133 my, a linear time scale is applied, based on Fig. 5.16 in ENAY et al. (1984), and plate 1 of GYGI & PERSOZ (1986).

In the western part of the Jura mountain range, the chronostratigraphy of the base of the Cretaceous is based on STEINHAUSER & CHAROLLAIS (1971) and CLAVEL et al. (1986); the "Urgonian" platform has been dated by CLAVEL et al. (1987). The Middle to Upper Berriasian, 131 to 128 my, comprises the Pierre Châtel, the Vions and the "Chambotte inférieure" formations.

The principles of dating in the Helvetic stratigraphic sequences have been exposed by FUNK (1985). Because of uncomplete fossil record, dating is often based on facies correlation. Numerous studies have shown facies diachronism in the Helvetics; so we possibly introduced errors up to 3 my.

For detailed references, see Table 1, parts A and B.

c) Depositional depth

The paleobathymetry has been estimated for the beginning and the end of deposition of each bed, and geohistory diagrams are presented here (Fig. 3) for a minimum and a maximum estimation of depositional depth. The main bathymetric evidences are the paleontological record, the petrographic composition and the sedimentary structures of the rocks. However, literature data are mostly insufficient to make sound evaluations of waterdepth, and therefore, somewhat simplistic assumptions have been made:

0 m water depth has been postulated for the deposition of the Triassic evaporites, for the "Purbeckian" rocks with pedogenesis (STRASSER 1987) and for the post Cretaceous to Eocene iron bearing "Bolus" and "Sidérolithique" formations. However, deposition above sea-level has also to be considered for continental sediments.

0 to 30 m water depth (sometimes 10 to 30 m) has been assumed for the high-energy zone, above the wave base. However, storm wave base may be considerably deeper in some cases (e.g. AIGNER 1985). Typical sedimentary units which have been ascribed to this depth are different oolitic limestones of the Middle Jurassic ("Grande Oolite" and "Hauptrogenstein"), the Oxfordian reef limestones of the northern and central Jura ("Rauracian" facies) and the biogenic or oolitic platform limestones of the Barremain to Lower Aptian "Urgonian" and "Schattenkalk" facies. Low energy back-reef facies of the top of the Upper Jurassic in the western and southern Jura ranges (BOLLIGER & BURRI 1970, BERNIER 1984) and most of the Middle to Late Eocene Helvetic algal and lagoonal limestones, sandstones and conglomerates (Subalpine Chains) have also been assigned to this depth.

20 to 50 or 80 m are depth estimations for some bioclastic and ammonite bearing sediment units, such as the Lower Cretaceous "Calcaire roux" in the western Jura and
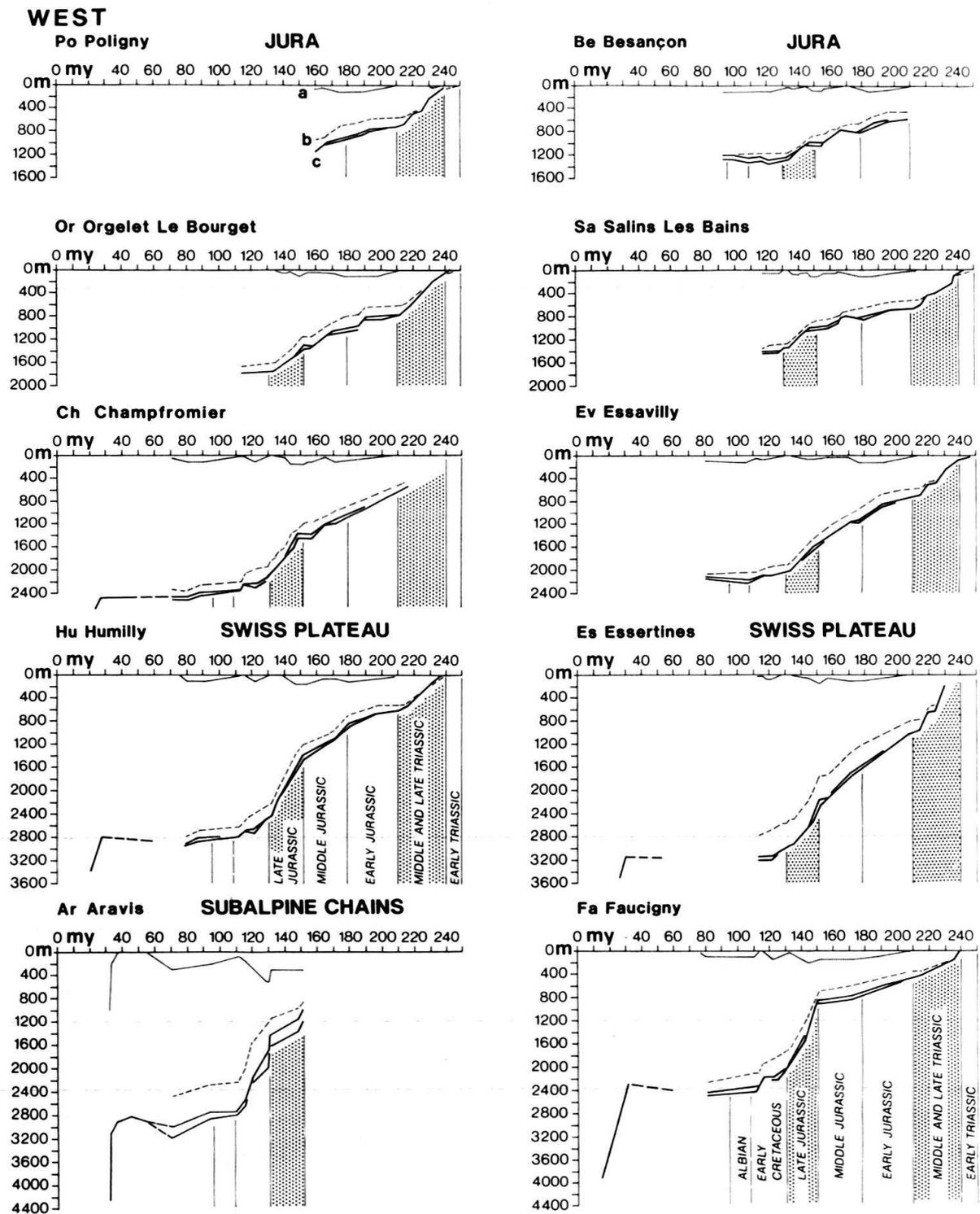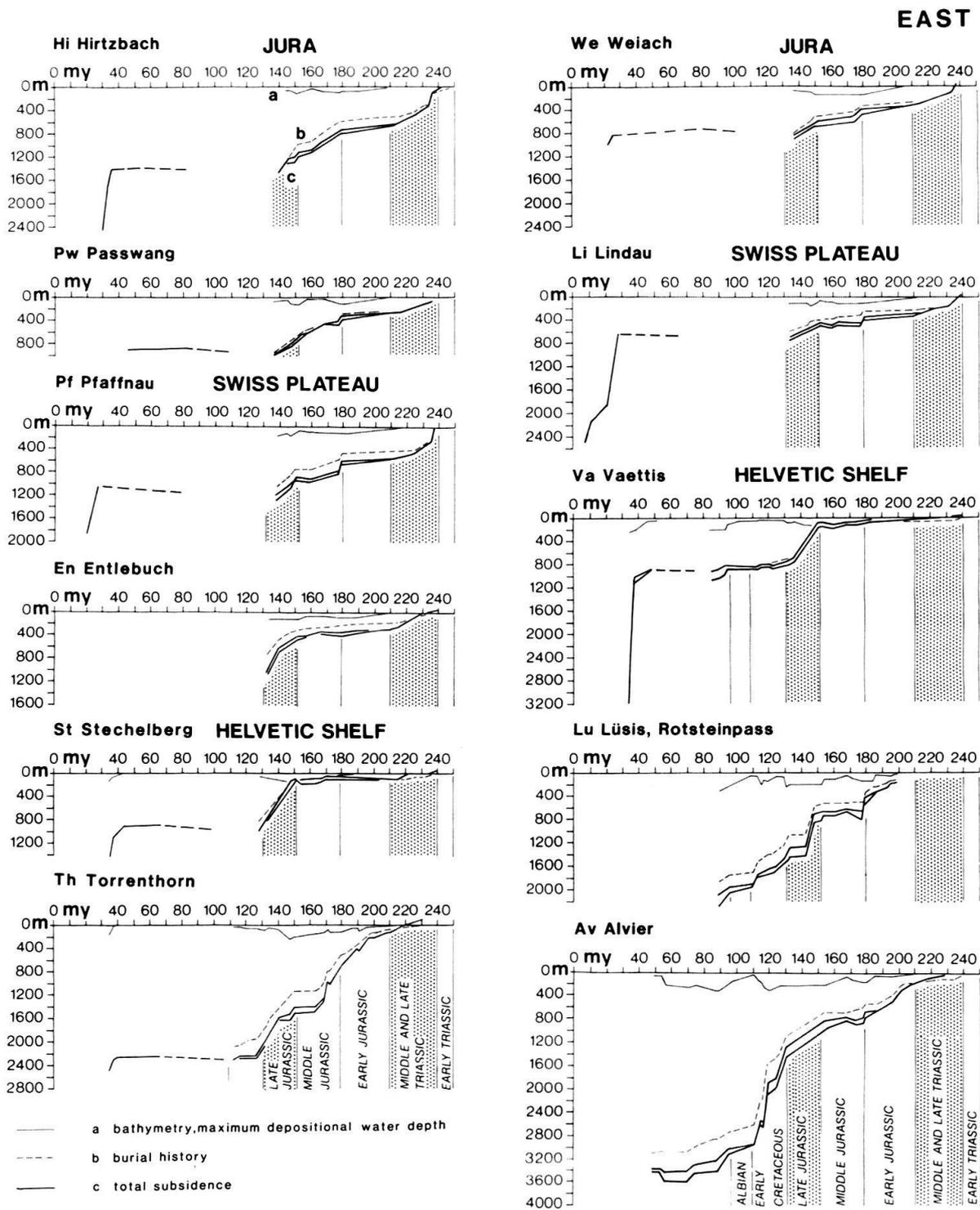
Fig. 3. Geohistory diagrams of the base of the Mesozoic sediments for some caracteristic sections of the Helvetic realm, the Swiss Plateau and the Jura (position of the sections: see Fig. 1 and 4).

a: maximum estimated depositional water depth.

b: burial history for actual thickness of sediments, corrected for tectonic deformation.

c: geohistory diagrams for maximum and minimum water depth, decompacted and stepwise recompacted sediment thickness and eustatic sea level fluctuations (HAQ et al. 1987, long term fluctuations).

Middle Jurassic crinoidal limestones in all considered areas. However, the depositional conditions of Mesozoic crinoidal limestones are still enigmatic and actualism is difficult to apply (see Roux et al. (1988) for analysis of deep-water crinoids). Most of the Liassic sediment units of the "Helvetic sandy facies zone" (e.g. Trümpy 1971) were also assigned to these depths.

50 to 100 or 150 m depositional depth are assumed for most ammonite bearing marly and shaly sediments, such as those of Middle to Late Liassic age in the Jura, below the Plateau and in the Helvetic nappes ("Dauphiné facies zone", Trümpy 1971) the Opalinus shales of the whole area, the Helvetic "Schilt" beds and the beds of "Argovian" facies (Oxfordian). For the latter, a detailed analysis based on ammonite faunas has been performed by Gygi & Persoz (1986) and Kugler (1987). A similar depth may be admitted for most of the Upper Jurassic Quinten limestones, for the Lower Cretaceous marls and pelites in the Jura mountains ("Marnes d'Hauterive") and in the Helvetics (Zementstein beds, Oehrli shales, Palfris marls) and for the siliceous limestone and similar lithologies of the Helvetic sections (Lüsis for instance). The Helvetic Globigerina shales have also been assigned to this depth.

Late Cretaceous water depth on the European borderland of the Tethys is very uncertain. 100 to 150 m have been suggested in the western Jura and a maximum of 300 m has been estimated for the Seewen limestone in the Helvetic realm (see also Oberhänsli-Langenegger 1978 for a deeper bathymetric interpretation).

Also, as the fossil record is sometimes poor in the Alps, a somewhat larger space between minimum and maximum depth of deposition has been introduced.

In any case, these data remain very hypothetical, especially for depths between the storm-wave base and the aragonite compensation depth, but they do not influence too much the general trends of the subsidence history diagrams (chapter 4).

d) Eustatic sea-level variations

The computed metric values used for the sea-level fluctuations are those of the long-term variations of Haq et al. (1987). Despite the critics of Burton et al. (1987), and as the general trends are in concordance with results found by other methods (e.g. Hallam 1978, Hays & Pitman 1973), the long-term changes give fairly sound values.

4. Mesozoic subsidence rates and regional subsidence phases

The subsidence rates (Fig. 5a to 5f) of the investigated area have been measured from the geohistory diagrams (Fig. 3 for characteristic sections) between 240 my Muschelkalk) and 96 my (end of the Albian), that means from the moment when the area was flooded by the sea, to the end of subsidence in most areas. The time of deposition of Buntsandstein (250 to 240 my) is not presented here as a consequence of bad dating and insufficient control of the elevation of the sedimentary realms with respect to sea level.

The analysis of the geohistory diagrams (Fig. 3) shows that every realm is affected by clearly defined phases of rapid subsidence seperated by periods of relaxation or of slower linear subsidence. The main phases in the subsidence history are:
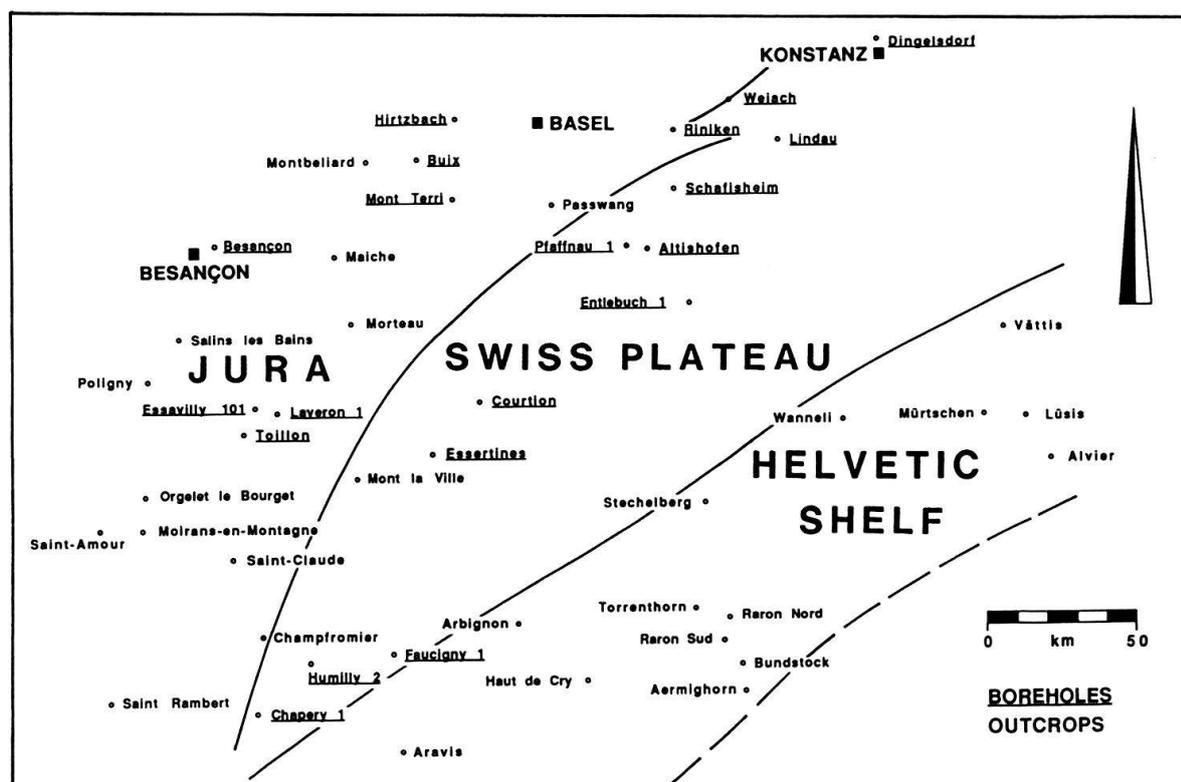
Fig. 4. Palinspastic position of the analysed sections.

a) *240 to 210 my (Middle and Late Triassic, Fig. 5a)*. – Fairly high subsidence rates are indicated in the ENE-WSW striking Jura realm, with maximum values of about 180 m/my in the salt bearing Keuper of the French Plateau Jura (St Amour, Laveron, Toillon, u.s.f.; the extreme value for Laveron 1 might be due to tectonic doubling of the evaporites). The starting point of the hyperbolic geohistory diagrams has most probably to be placed at the time of continental sedimentation in the Early Triassic, or even in the Permian or the Carboniferous. Relaxation begins at the end of the Triassic. The Helvetic realm remains stable with rates of normally less than 10 m/my.

b) *210 to 179 my (Early Jurassic, Fig. 5b)*. – The subsidence rates are abruptly diminishing in the Jura, the Plateau and the northernmost part of the Helvetic realm to values less than 10 m/my and in places to zero. Higher rates (up to 25 m/my) subsist in the Geneva area, in the southernmost part of the Helvetic realm and in the western Helvetic shelf in general. Therefore, relaxation in the major part of the study area contrasts with a more or less rapid subsidence in the west and the south.

c) *179 to 152 my (Middle Jurassic, Fig. 5c)*. – Rapid subsidence in the western Helvetic realm is confirmed during early and middle Middle Jurassic with rates up to about 40 m/my (Torrenthorn, Haut de Cry), followed by relaxation during late Middle Jurassic. On the top of the Aiguilles Rouges basement, inversion by the tilting of a basement block and erosion may be suspected during late Early Jurassic and early Middle Jurassic (BADOUX 1972). The eastern and southernmost Helvetic realms remain fairly stable with low rates. The basement of the Jura subsides in an irregular
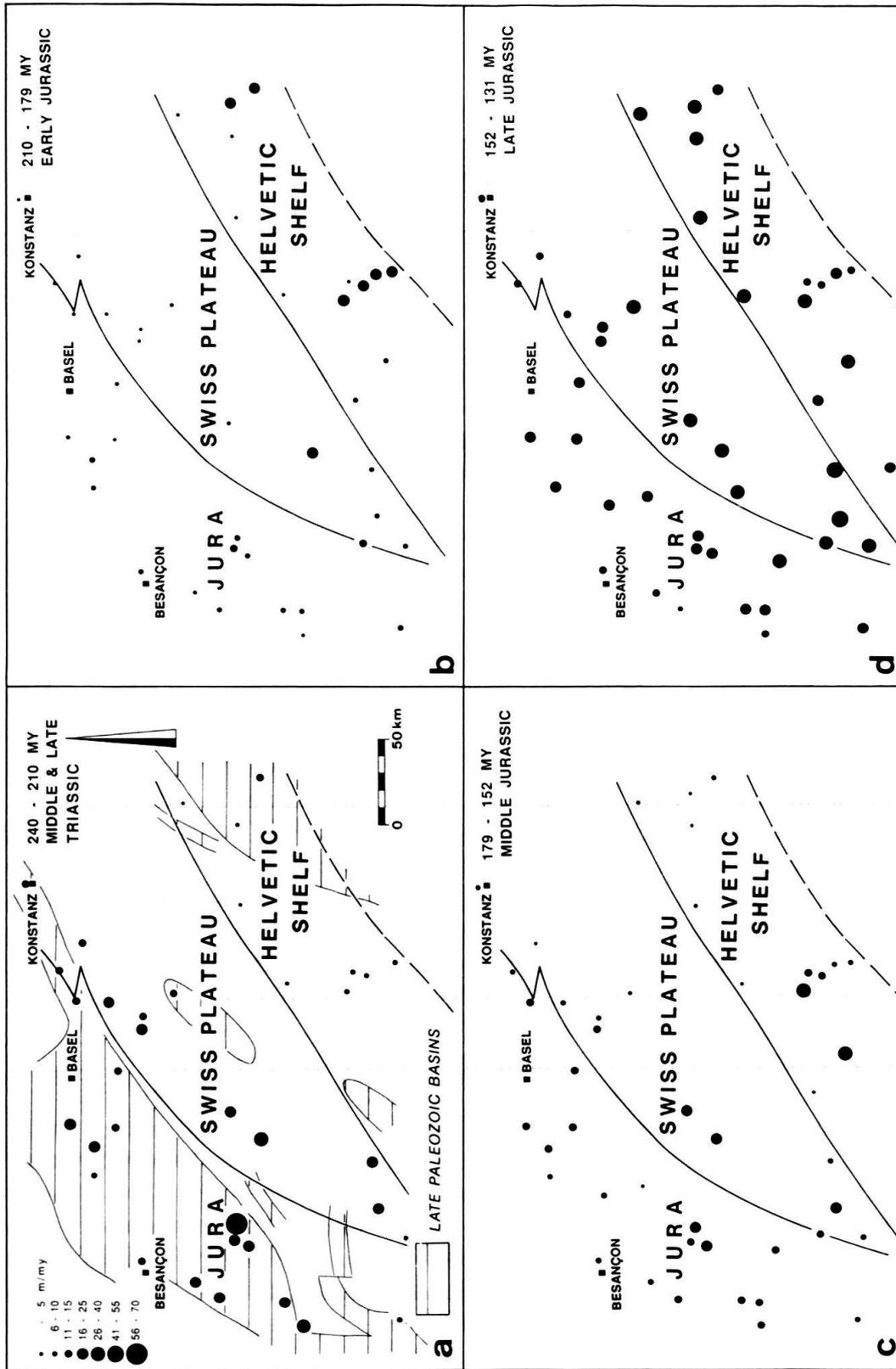
Fig. 5a–d. Maps of mean subsidence rates in the Helvetic realm, the Swiss Plateau and the Jura from the Middle Triassic (Muschelkalk) to the Late Jurassic. (a: main Late Paleozoic basins after LEMCKE 1961, ZIEGLER 1982, DEBEGLIA & GABLE 1984, MÜLLER et al. 1984, KLINGELE & MÜLLER 1987; actual position for the Jura and the Swiss Plateau, palinspastic restauration for the Helvetic realm).
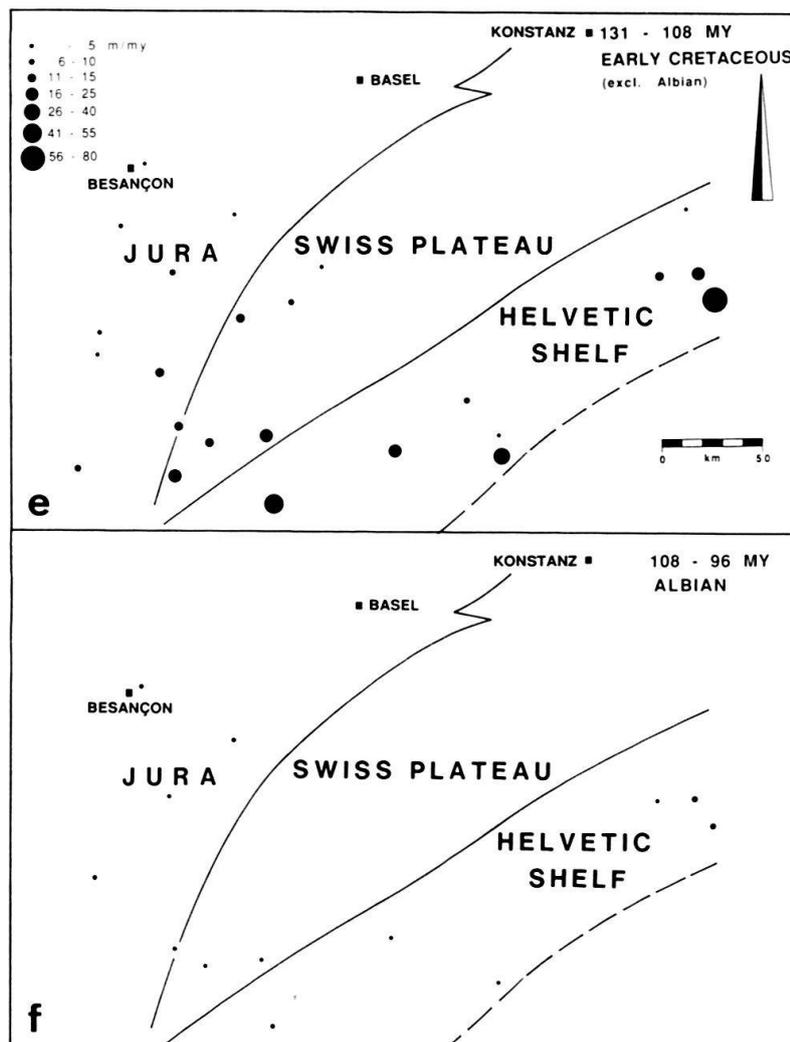
Fig. 5e+f. Maps of mean subsidence rates in the Helvetic realm, the Swiss Plateau and the Jura during the Cretaceous.

pattern with rates of 10 to 40 m/my, indicating a moderate but real acceleration with respect to the Early Jurassic.

d) *152 to 131 my (Late Jurassic, Fig. 5d).* – The whole shelf, from the Helvetics in the south to the Jura in the north, subsides abruptly, with short-time values as high as 70 m/my on the top of the basement of the Aare massif and in the eastern Helvetic realm, and 20 to 40 m/my in the southern part of the Jura. Subsidence is somewhat attenuated with respect to these areas in the western part (except for Torrenthorn and Haut de Cry sections) and in the southernmost part of the Helvetic realm. To the north, the area of rapid subsidence is limited to the area of Dijon.

e) *131 to 108 my (Base of Cretaceous to Late Aptian, Fig. 5e).* – Subsidence is diminishing to rates lower than 10 m/my, in the northern part of the western Jura and in the realm of the Aare massif, and to rates lower than 20 m/my in the northern part of the Helvetic realm. In the south-western Jura, values of 20 m/my at the beginning slow down to 10 to 15 m/my at the end. In the southernmost Helvetic realm (Aravis, Aer-

mighorn, Alvier), a further subsidence phase starts at 120 my with subsidence rates of up to 100 m/my. No sediments of this age are preserved in the eastern Jura and Swiss Plateau and the end of the Mesozoic subsidence history is thereby unknown in this area.

f) *108 to 96 my (Albian, Fig. 5f)*. –The subsidence due to passive continental margin evolution is terminated. However, local subsidence may still occur, but orogenic crustal flexure may already be of some importance. Therefore, the Late Cretaceous and Tertiary vertical movements of the crust will be discussed in a forthcoming paper.

In addition to these regional subsidence phases, many events of local importance which are not discussed here can be deduced from the computed geohistory diagrams.

5. Subsidence mechanisms

In this work, a more or less "statistical" approach based on available vertical summary logs has been chosen in order to determine the subsidence history. However, this approach is but partly appropriate to discuss the mechanisms of subsidence, and other types of informations are needed:
— data on the geology of the pre-Mesozoic basement;
— data on the three-dimensional geometry of the Mesozoic and Cenozoic sediments;
— data on synsedimentary faults and scarps;
— data on the thermal history of the sedimentary basins.

In addition to this, most parameters used here are susceptible to be modified by the results of further research work and the following discussion is but of qualitative value.

a) The *Triassic subsidence phase* in the Jura basin has to be examined in the context of the Germanic Triassic basins, rather than in relationship with the opening of the Tethys. In the entire Jura basin, an evident correlation exists between the existence of Permian or Carboniferous late- and post-variscan basins and the Triassic basins (Fig. 5a) (see also BRUNET & LE PICHON 1982 for the Paris Basin). A very rapid decrease of the heatflow in the Late Paleozoic has been deduced in Northern Switzerland from vitrinite reflectance data by KEMPTER (1987) and rifting and strike-slip tectonics has been postulated by LAUBSCHER (1987). Synsedimentary faults have been proposed to exist in the Rhine valley of the vicinity of Basel, where rapid lateral variations of thickness of salt layers have been observed (HAUBER 1971). These data may indicate a Late Variscan to Triassic strike-slip and rifting phase, followed by thermal relaxation as in the Paris basin. However, more tectonic and paleogeothermal data are required to improove or reject this hypothesis. In contrast to this situation, no significant correlation can be established in the Helvetic realm between Triassic subsidence and the presence of Permian "Verrucano" basins with the exception of profile "Alvier" which lies possibly on top of the "center" of this basin.

b) The locally high *Early Jurassic subsidence* rates in the Helvetics are commonly related to extensional tectonics inducing fault-controled basins on the southern part of the Aare and Aiguilles Rouges massifs (TRÜMPY 1949, BAER 1959, DOLVIO 1982, HUGGENBERGER 1985). Although paleofaults are in places fairly well documented, there is, in fact, little field evidence for most areas where they have been deduced. Anyway, the obliquity between the Liassic facies zones ("Daupiné" shaly facies and

"Helvetic" sandy facies) and the Alpine structures (e.g. TRÜMPY 1949, 1971) must be stressed and pull-apart seems a sound explanation for the orientation of these basins. Also, in most parts of the study area, the Liassic corresponds to a time of relaxation.

c) The *Middle Jurassic subsidence phase* in the Jura basins follows the regional trend of the Triassic subsidence; a remobilization of ancient structures at the beginning of Middle Jurassic seems probable. In the Helvetic realm, some synsedimentary faults may exist which seem to be of local or regional significance (GÜNZLER-SEIFFERT 1941, TRÜMPY 1949, BAER 1959, for a discussion see LEMOINE & TRÜMPY 1987).

d) The *Late Jurassic subsidence phase* is the first phase with clearly WSW-ENE (that means parallel to the Alpine Tethys) striking zones of equal subsidence but not of equal facies, and a roughly north-south polarity of increasing subsidence. Extensional structures with small-scale tilted blocks have been reported from the Subalpine Chains by DETRAZ (1988). On the other hand, no tectonic evidence for crustal extension is known, neither in the Jura, nor in the Swiss Plateau, nor in wide areas of the Helvetic realm. Sedimentary evidence however indicates differential subsidence in the Oxfordian of the Jura (KUGLER 1987) as well as in some parts of the Helvetic area (ROD 1937, DIEGEL 1973) for Late Tithonian age.

As a conclusion, one is tempted to see a link between basement subsidence of the marginal shelves and the spreading of the Tethyan basins. It may correspond, for instance, to low angle faulting, extending from the Jura basin to the Helvetic shelf. However, as such faults have not been reported up to now, other density changing events in the lithosphere as a consequence of Late Jurassic ocean floor formation in the Tethys, have also to be considered: indeed, the possibility of plastic extension of the lower crust in a certain distance from the centre of brittle extension of the upper crust has been demonstrated by small-scale models to be a possible subsidence mechanism (ALLEMAND et al. 1989).

e) For the *Early Cretaceous subsidence phase* at the Helvetic margin, extensional features have been mentioned by GÜNZLER-SEIFFERT (1952), SCHINDLER (1959), ISCHI (1978) and STRASSER (1982). Clockwise north-south strike-slip movement are indicated by synsedimentary cracks at the top of the Schrattenkalk Formation (GERBER & OUWEHAND 1988). These events coincide with the first eoalpine orogenic events in the Eastern Alps, in the south-Penninic and probably north-Penninic basins, and with the time of deposition of thick terrigenous Flysch and Bündnerschiefer deposits (FRANK et al. 1987, WILDI 1988, cum biblio). Therefore, the possibility of a link between the subsidence phase and the first inversion movements in the Austroalpine and Penninic realms has to be considered.

6. Discussion and conclusions

The analysis of the subsidence history of the European marginal shelfs exposed here above is based on data of very unequal origin and reliability: fairly safe thickness measurements contrast with missing informations on compaction and cementation, with schematic sedimentation depth evaluations, eustatic data and palinspastic reconstruction which are still a source of discussion. Also, radiometric time scales will certainly be revised in the future which could change the different sedimentation rates (FUNK 1985, Fig. 4).

Despite these systemtic errors, the numerical analysis indicates some interesting trends on the subsidence history and on the mechanisms:

— the subsidence phases of Triassic and Middle Jurassic age are most pronounced in the Jura basin, where Late Paleozoic sediments on the top of the basement may indicate a link between Mesozoic and Late Variscan events within the European lithosphere, which are in no relation with Tethyan rifting;

— the Liassic subsidence which has previously been considered to be the most charcteristic in the Helvetics is of rather local than regional importance in most places. If the Liassic marine transgression of the Helvetic shelf can partly be related to extensional tectonics with fault-bounded subsiding basins, other processes such as eustatic sealevel fluctuations act as a major control on the sedimentology.

— the further subsidence phases of Late Jurassic and Early Cretaceous age are most probably linked to the last extension and the first inversion of the Alpine Tethys. However, only little fault tectonics is known for these phases.

Acknowledgments

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