CHAPTER 8

LILLEVATN MEMBER

8.1 Introduction

The Lillevatn Member, the lower part of the Stappogiedde Formation, has been informally subdivided into a fine grained lower submember and coarse grained upper submember. As mentioned in Chapter 2, the lower submember is thick, up to about 50 m, south of Sjursjok, and thin, about 3-6 m, north of Sjursjok.

The upper submember is an influx of coarse sediment between the fine sediments of the lower submember and of the Innerelv Member. In contrast to the lower submember it is of remarkably constant thickness, about 40 m over the whole area studied, from Austerelva, south of Laksefjord, to Leirpollen, a distance of about 75 km.

8.2 Thick Lower Submember

8.2.1 Description

The lower contact of the submember is rarely exposed, but is gradational when observed. In the lower part the submember consists of grey parallel-laminated silty mudstone, which alternates with tillite at the contact with the Mortensnes Tillite (see Chapter 7). Upwards, the grain size gradually increases from silty mudstone to muddy siltstone, and silty very fine sandstone. Concomitantly, lamination thickens upwards, and near the top, small-scale ripple cross-lamination occurs. In the top few metres, thin, lenticular, fine and medium grained sandstones occasionally occur. These are usually sharp based; they may be massive, parallel-laminated or cross-bedded, and are micaceous and feldspathic. The contact with the upper submember usually appears sharp, but at a few contacts steep-sided erosion surfaces cut as much as 1 m into the lower submember.

North of Vestertana, over a distance of about 2-3 km, the thick submember thins from about 35 to 5 m (fig. 81). In this transitional zone the lower contact of the lower submember was not observed.

Palaeocurrent evidence was not obtained from this unit.

8.2.2 Interpretation

A large-scale coarsening upwards sequence from mud to sand, concomitant with increasing agitation upwards indicates gradual shallowing, and is similar to sections through the marine part of the Rhone Delta (Oomkens, 1970). Such a sequence probably formed by the progradation of a shoreline, either marine dominated, or deltaic, into relatively deep, quiet water (Reading, 1971).

8.3 Thin Lower Submember

8.3.1 Description

This submember consists of sandy or silty grey mudstone, mostly parallel-laminated, with occasional ripple cross-lamination. Where the base is seen, the submember sharply overlies the conglomerate bed at the top of the Mortensnes Tillite (figs. 62, 82). North of Guoholmen, on the Digermul Peninsula (fig. 69), ferruginous mudstone, with pyritiferous coarse sandstone lenses occurs at the base of the thin lower submember. The section at Lavvonjargga (fig. 82) is unique because of the gradual coarsening upwards from mudstone to sandstone. The contact with the overlying upper submember is still sharp.

Intercalated with the mudstone typical of the submember are various types of sandstones, mostly medium grained. These may be rippled, massive, or parallel-laminated, and may have sharp or gradational contacts with the mudstone. An unusual

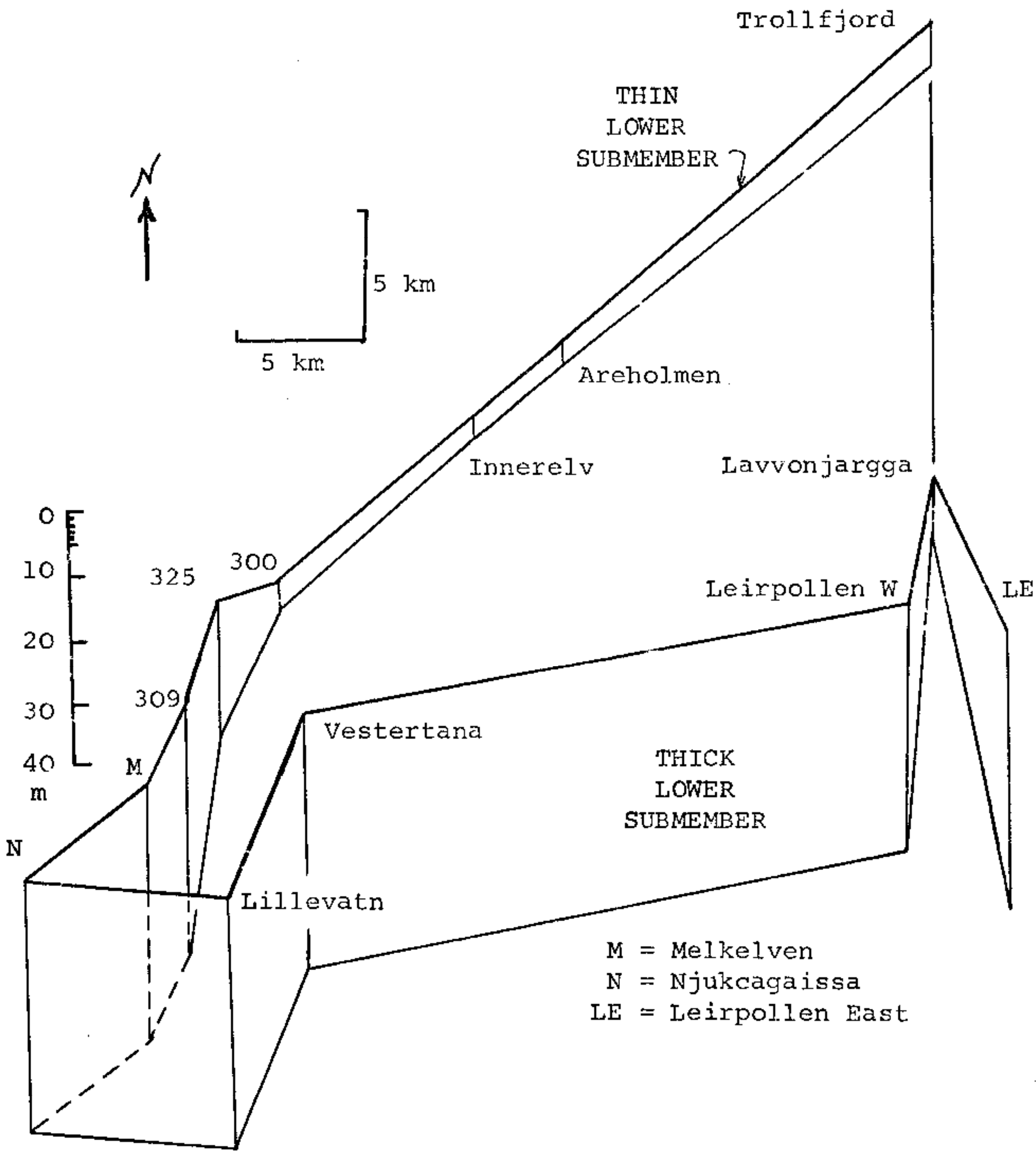


Figure 81. Variation in thickness of the Lower Submember, Lillevatn Member. Datum is upper surface.

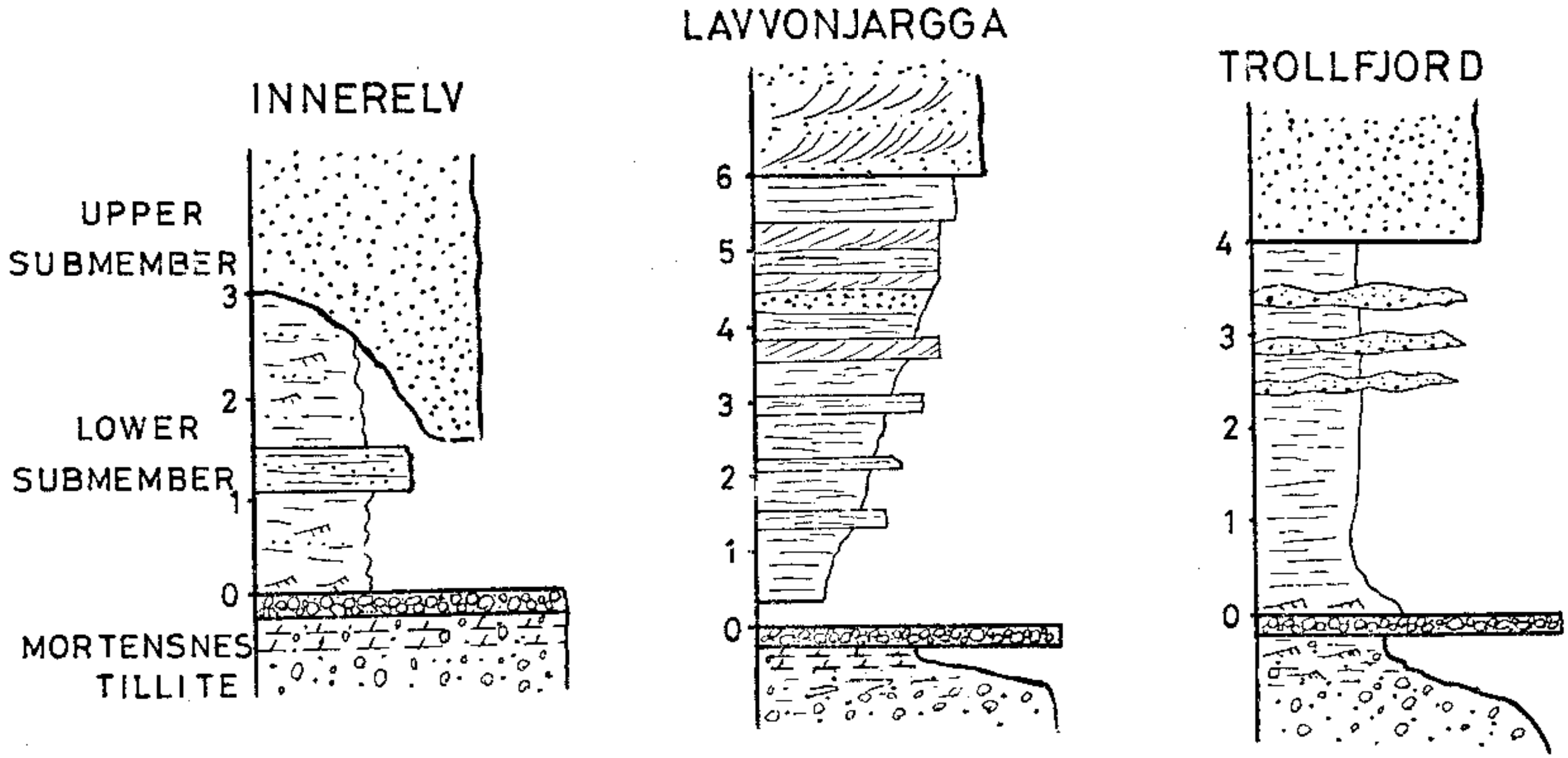


Figure 82. Sections through the thin lower submember, Lillevatn Member.
Thickness in metres.

occurrence is the lenticular graded beds at Trollfjord (fig. 82).

8.3.2 Interpretation

The general occurrence of ripple cross-lamination in the thin lower submember around the southern part of Tanafjord suggests that deposition was in relatively shallow water compared to the basal part of the thick lower submember. The presence of ripples only at the base of the section at Trollfjord suggests that conditions deepened rapidly. The ferruginous mudstone at Guoholmen suggests very slow deposition, during which time the bottom was enriched in iron.

The occurrence of the conglomerate bed of the Mortensnes Tillite immediately below the submember suggests that conditions changed rapidly from high to low energy. Such a change may have been caused by a rapid relative rise in sea-level. It thus seems likely that the thin lower submember was deposited over a relatively elevated area compared to the thick submember, the difference in relief causing the observed change in thickness.

8.4 Upper Submember

8.4.1 Introduction

In contrast to the lower submember, the upper submember is a complex assemblage of varied lithologies which are classified into three groups of facies. Two groups of facies A and B occur in the lower part of the member, while facies C occurs in the upper part. Facies A consists mostly of sandstone with some fine conglomerate. Mudstone is rare. Facies B is mostly mudstone, in parts with intercalated sandstone beds. Facies C consists of varied rocks, including sandstone and mudstone.

8.4.2 Environmental Context

To facilitate the presentation of the detailed interpretation of the facies, the broad origin of the upper submember is first considered. As the thick lower submember represents the progradation of a shoreline, the upper submember is either a coastal sand, or a fluvio-deltaic deposit. Three observations suggest that the latter origin is correct:

1) Sandstone in facies A are typically poorly sorted, with a considerable proportion of matrix, and occasional fine conglomerate.

2) Compositionally, the sandstones are feldspathic and micaceous, and the mudstones are micaceous.

3) The coarse sandstones, and the mudstones in the lower part of the submember may occur separated from each other rather than interbedded (fig. 83)*.

With regard to texture and composition, the rocks resemble closely fluvial deposits, but contrast with typical shallow marine deposits (e.g. Reading, 1970). In a fluvial context, the coarse deposits (facies A) formed in channels, while the fine deposits (facies B) formed on the floodplain. Facies C in the upper part of the upper submember, passes up into the marine shales of the Innerelv Member (Banks, 1971) and was probably deposited in coastal environments during a transgression (Reading and Walker, 1966).

8.4.3 Facies A, Channel Deposits

<u>Description</u>

Two channel facies can be recognised in most sections (fig. 83). Although both facies are basically medium bedded sandstones, the sorting and grain size of the facies contrast.

Facies A1

Facies A1 consists of medium grained, moderately sorted, subarkosic sandstone which usually occurs in medium beds, 10-40 cm thick. Beds are either parallel-sided or slightly lenticular. White grey sandstones with a silica cement predominate, but in the upper parts of some sections grey-black, haematite cemented sandstones occur.

Parallel lamination is the most common structure, but massive, and trough and planar cross-bedded varieties are also frequent (Pl.191).

This facies often occurs in the basal few metres of the upper submember, and is followed by facies A2. The facies also occurs in the higher parts of many sections, 10-20 m from the base of the upper submember (fig. 83).

Facies A2

This facies consists of very poorly sorted coarse, subarkosic sandstone with granule conglomerate and a muddy micaceous matrix in many beds. Pebbles up to 3.5 cm long were seen at Melkelven (fig. 83), but an average size of 2-4 mm is typical. The matrix imparts a dark grey colour to some beds.

Bedding is medium to thick, and parallel-sided or lenticular; rare scours are seen within the facies (Pl.192). Massive beds predominate, but trough cross-bedding occurs, and in one outcrop very faint, irregular subhorizontal lamination was noted.

The facies is restricted to the lower 10 m of the submember where it occurs erosively or apparently sharply on facies A1, or occasionally at the base of the section on the lower submember.

Facies A: Interpretation

Within a continuum of forms, channels are distinguished as straight, meandering and braided (Leopold and Wolman, 1957).

Although many explanations have been offered for the development of the different patterns (see Schumm and Kahn, 1971), the importance of the relationships between sediment load, flow characteristics and climate has been shown by Maddock (1969). In relative terms, meandering channels typically have fine bed load, moderate slope and stable banks, while braided channels have coarse bed load, high slopes, and unstable banks (Table 25). Braided channels also appear to form where the extremes in discharge are pronounced. The properties of straight channels are less well documented, but it has recently been shown that straight channels are associated with very low slopes, other factors being equal (Schumm and Kahn, 1972). In addition, suspended load appears to stabilize banks, causing channel erosion, and ultimately increasing sinuosity (Schumm and Kahn, 1972).

The contrasting deposits of meandering and braided stream deposits are well documented. The overall fining up of a meandering channel deposit is a result of large-scale helicoidal flow and lateral migration of the channel and point bar (Allen, 1970) which may compose 80-90% of a meandering river flood plain (Leopold and Wolman, 1957). The important processes in braided channels appear to be the filling of minor channels (Doeglas, 1962; Williams and Rust, 1969) and the migration of flat longitudinal and transverse bars (Collinson, 1970; Smith, 1971). The channel deposits do not fine up, and scour-and-fill structure and cross-stratification are abundant. Considering the absence of fining upwards, and of critical sedimentary structures in the present deposits, neither the meandering nor braided channel pattern appears to be applicable to facies A.

Deposits of streams with straight channels are not well known. Their morphology has been discussed by Leopold and Wolman (1957), and Leopold, Wolman and Miller (1964). The only description of natural straight channel deposits familiar

Channel type:	Meandering	Braided	Straight
PROPERTIES			
Channel Sinuosity	High	Low	Low
Thalweg Sinuosity	High	High	High
Slope	Moderate	High	Low
Typical Discharge Variation	Low	High	Variable ?High
Bed Load	Fine	Coarse, variable	Variable, coarse
Bed/Susp.Load	Low	High	?mod.-?high
Banks	Stable	Unstable	?moderate
W/D	Moderate	High	Moderate
Large scale bed resistance	Point bars, meanders	Braid bars, High W/D	Side bars, thalweg meander
X-section profile	Asymmetric	Irregular	Asymmetric
Bed relief	Low	High	Low
Deposition sites	Point bar, thalweg	Mid-channel bars, channels	Side bar, channel
Channel/Overbank thickness	4/5	c.1	??
Sedimentary Structures	Trough x-strat, parallel lam. Ripple x-lam.	scour & fill, x-strat.	Trough x-strat, Large scale foresets.
Grain size trend within channel	Fines up gradually	Irregular	Irregular
Large Structure	Epsilon cross-strat.	Interlocking channels	???

Table 25. Characteristics of fluvial deposits. See text for references.

to the present author is that of McGowan and Garner (1970) of the relatively coarse grained Colorado and Amite Rivers. Flume experiments have been carried out by Maddock (1969) and by Schumm and Kahn, (1971, 1972). Models of straight channel deposition have been proposed by Allen (1965) and Moody-Stuart (1966).

Straight channels are characterised by alternating side bars with adjacent pools, riffles or crossovers, and a meandering thalweg (fig. 84) (Leopold, Wolman and Miller, 1964). These features are analogous in form to the point bar, thalweg and pool and riffles of a meandering channel (Leopold and Wolman, 1957; Tinkler, 1970; Keller, 1971), and suggests that there is some continuity of processes between the two types.

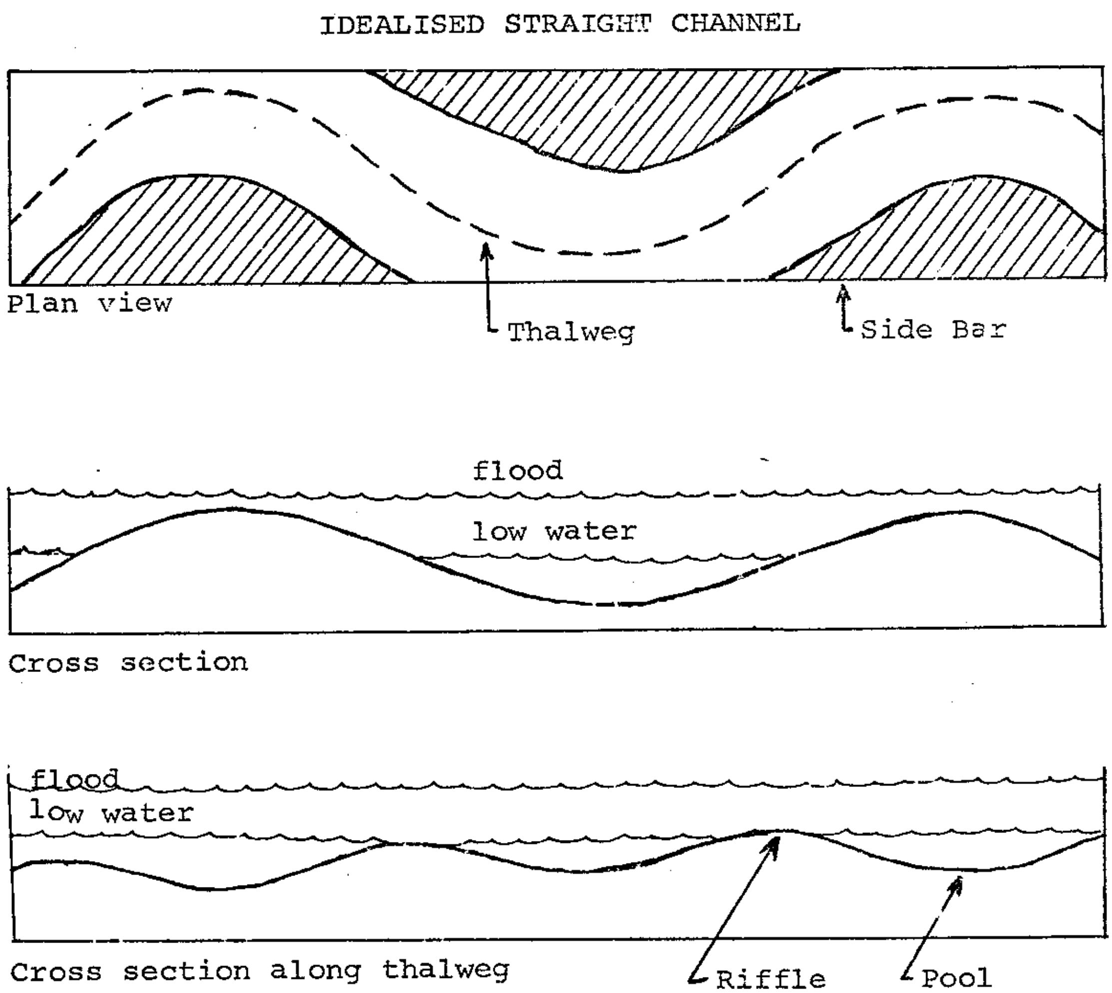


Figure 84. Morphology of straight channel with meandering thalweg.

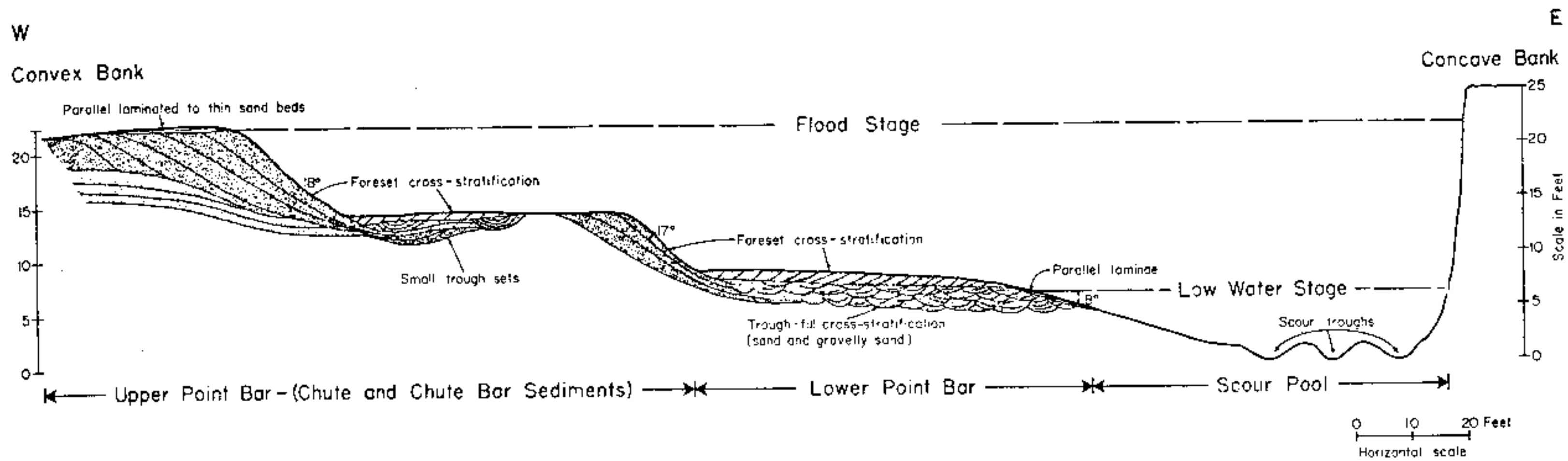


Figure 85

Profile and stratification from the scour pool to the crest of the highest chute bar, Amite River, Louisiana. (From McGowen and Garner, 1970, fig. 4).

The coarse-grained point bar described by McGowen and Garner (1970) from the Colorado River consists of two important parts (fig. 85): the lower point bar which is sandy and includes trough and foreset (planar) cross-stratification, and parallel lamination, and the upper point bar which is gravelly, and usually consists of large scale foreset stratification.

Topographically below the point bar is the scour pool (fig. 85) which is composed of pebbly coarse sand and may be trough cross-stratified, and above the point bar is the flood plain which, adjacent to the channel, consists of cross-stratified and parallel-laminated gravel and sand near the channel. The point bar accretes mainly in a downstream direction, during flood conditions, producing a vertical sequence with an irregular grain size trend, and characteristic sedimentary structures (fig. 86).

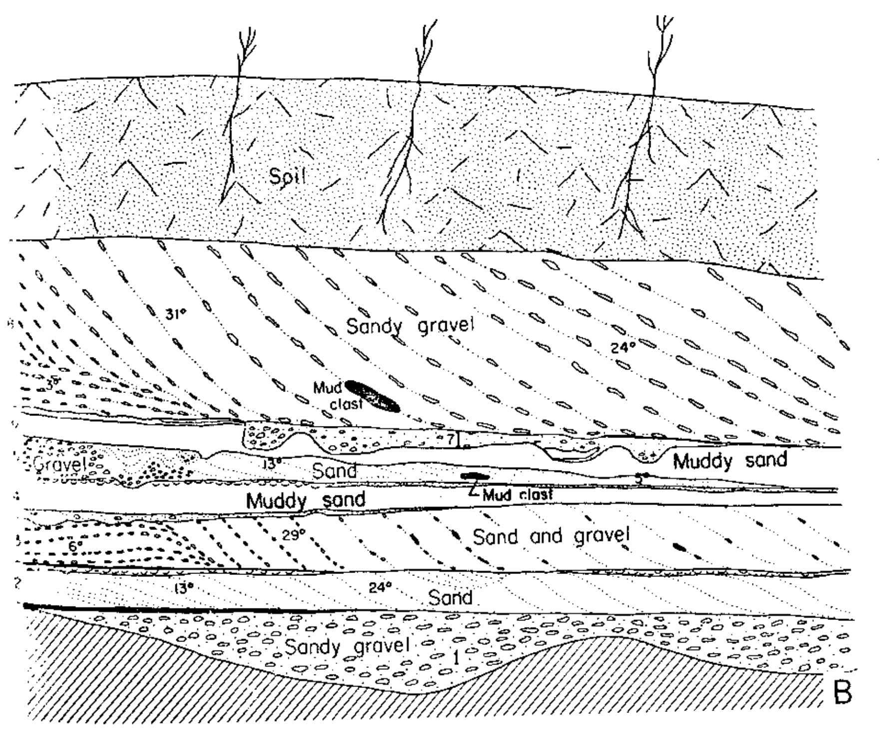


Figure 86. Vertical succession of stratification types in Pleistocene deposits, Travis County, Texas. Channel lag (unit 1), lower point bar or crossover (units 2-7), and chute bar (unit 8). (From McGowen and Garner, 1970, fig. 21).

The lower point bar described by McGowan and Garner has large gravel veneered transverse bars with intervening scour troughs. Trough fill cross-stratification was 5-15 cm deep on the Amite River, and 60-90 cm deep on the Colorado River.

The upper point bar typically has chutes and chute bars. Chutes are sinuous, linear scours aligned parallel to the main channel. They vary from over 1 to 5 m deep, at least 5 m wide, and around 100 m long. The floor is gravel covered but during the falling stage a sand-to-mud unit may be deposited within the chute (McGowen and Garner, 1970, p.86) (see also Harms et al., 1963). Chute bars, lobate features that form downstream from chute bars, accrete laterally and vertically forming large-scale foreset cross-stratification during upper flow regime conditions of extreme flood. Bar height is usually over 1 m to about 3 m. Chute and chute bar topography varies considerably and in some cases is very ill-defined (McGowen and Garner, 1970, p.91).

Side bars observed in rivers by Maddock (1969) were interpreted as flood formed, and which became emergent during low water. In flume experiments, side bars formed under water (see Schumm and Kahn, 1971), a condition analogous to flooding in natural streams. Thus, side bars are active during flood stage and seem to be associated with rapid flow and shallow depths.

During low water, the emergent parts may be modified by gullying, while bed load transport may take place in the thelwag. Such transport was observed in crossovers of the Colorado River (McGowen and Garner, 1970, p.79).

On the basis of the processes discussed by McGowen and Garner for coarse grained point bars, a model to explain the channel facies of the Lillevatn Member can be devised (fig. 87). This model attempts to explain the absence of the characteristic sedimentary structures they observed as well as the similarities between the deposits. According to this model, facies A_1 is

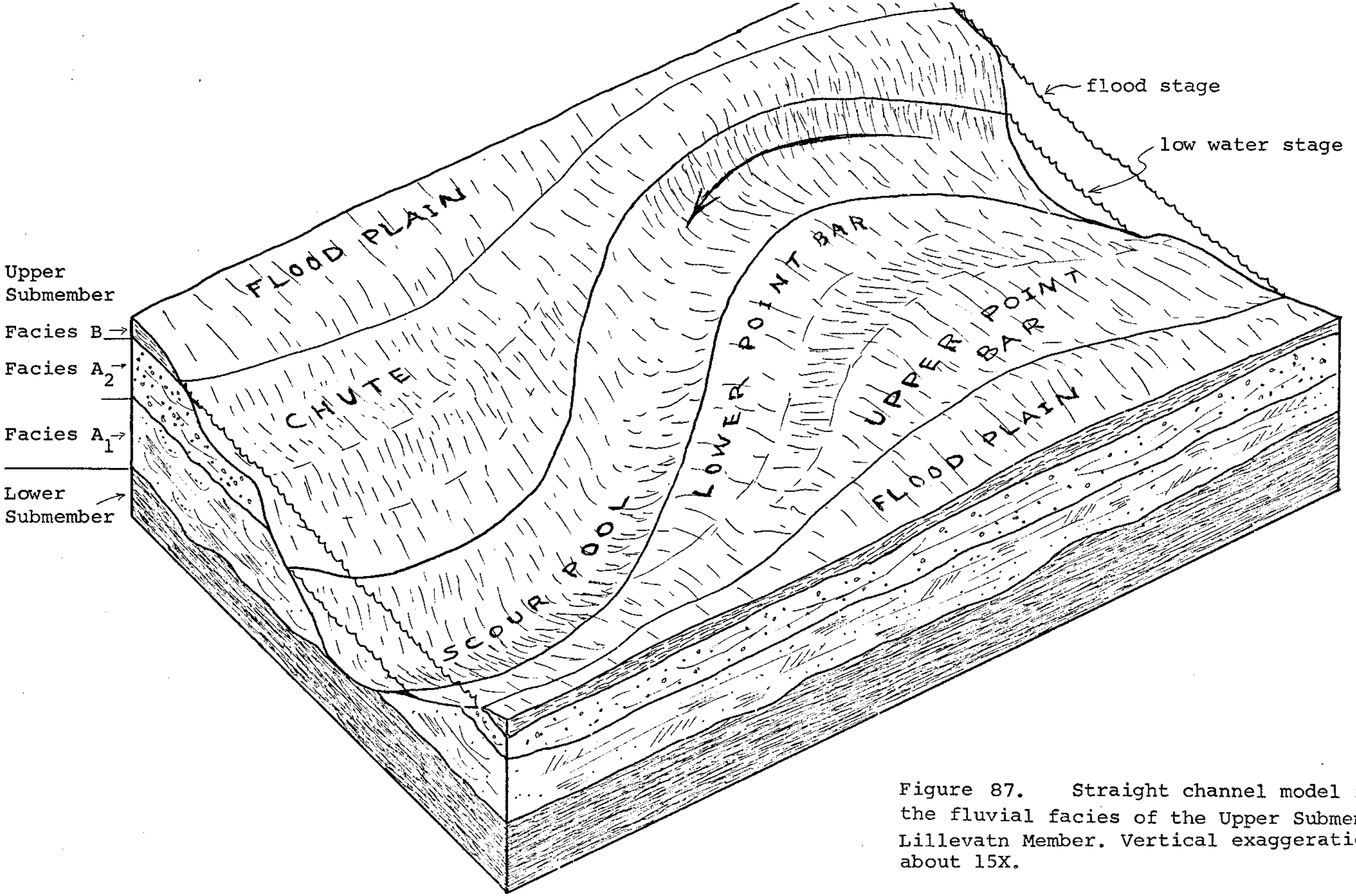


Figure 87. Straight channel model for the fluvial facies of the Upper Submember Lillevatn Member. Vertical exaggeration about 15X.

the lower point bar pool deposit formed in the upper part of the regime conditions, and in the lower part of the upper flow regime. These conditions may have prevailed during flooding associated with greater depths, or during the low stages with shallow and low velocity flow.

Facies A2 is the upper point bar deposit. The coarse grain and very poor sorting indicate strong flow with both coarse and fine material in suspension, suggesting rapid flow conditions at shallow depths, during flood. This may account for the prevalence of massive and parallel-laminated beds in facies A2. The one example of irregular lamination in facies A2 is similar to that described by Middleton (1967) in antidune deposits studied in a flume.

That facies A1 often comes at the base of a channel sandstone, and is followed by facies A2 agrees with the vertical relationship of the two facies in the model, and suggests that the upper point bar accreted forward and vertically over lower point bar-scour pool deposits.

The absence of large scale foreset cross-stratification in the upper point bar deposits of the Lillevatn Member suggests that well developed chutes and chute bars did not form. Instead, the relief on the bar may have been low, and gradients gentle, without a separation eddy forming in the lee of a chute bar. The presence of some chutes was attributed by McGowen and Garner (1970, p.85) to scouring in the lee of uprooted trees. Other processes for the initiation of chutes were not mentioned, although the presence of coarse gravel may serve to stabilize certain areas as bars. In any case, large scours within the facies A2 units may represent subsequently filled chutes on the upper point bar.

Models of straight channels proposed by Allen (1965, p.164-5) and Moody-Stuart (1966) deal with large scale properties rather than the detailed internal structure. According to Allen, the channels migrate laterally over the flood plain, while according to Moody-Stuart they move laterally by avulsion. Both authors agree that a fining upwards sequence within the bed load deposits would form. This is not explained by Allen, but according to Moody-Stuart, is related to gradual aggradation causing the channel to widen and shallow producing a waning flow suite of structures. Although such a suite was not formed in the sediments of the Lillevatn Member, the sharp junction with the overlying flood plain deposits suggests that avulsion, rather than lateral migration may have been an important process of lateral channel movement. However, the occurrence of facies A1 at the top of some channel deposits (section 5, fig. 84) may indicate gradual shallowing, or lateral migration.

If the channel facies of the Lillevatn Member were formed in straight channels with meandering thalwegs, then it can be speculated as to why braided streams, or meandering streams did not form.

In experiments on the relationship between slope and channel pattern, Schumm and Kahn (1971, 1972) showed the presence of two threshold slopes, dividing the range of slopes into three: less than .2%, .2% to 1.4-1.5%, and above 1.5%. At low slope straight channels with straight thalwegs formed, at moderate slopes, straight channels with meandering thalwegs, and at high slope, braided pattern. A meandering channel was not stable with the coarse, poorly sorted sand used. The experiment seems to indicate to the author that with non-cohesive banks a stable meandering channel cannot form. The fact that a meandering thalweg formed indicates that a "meandering tendency" was present in the flow conditions. A meandering stream may not have formed because the channel banks were not cohesive enough for the discharges and slopes involved. Sufficient fine load to cause meandering of the thalweg must have been present. The fine-grained floodplain deposits (facies B) shows that fine suspended load was being carried by the streams.

For the load and discharge of the upper submember facies A channels, the slope was low enough to prevent the development of meandering channels. A possible reason for this is discussed below.

8.4.4 Facies B, Flood Plain Deposits

Description

Facies B forms units composed predominantly of mudstone up to 20 m thick. The sandstone content may range from almost none to almost 100% of the total thickness. Units tend to rest sharply on facies A and are overlain by facies A sandstone or facies C sandstone and siltstone (fig. 85). Two floodplain deposits are considered as parts of a continuum including mudstone (facies B1) and sandstone interbeds (facies B2).

Facies B1

Facies B1 consists of dark grey, brown weathering rippled and finely parallel-laminated, silty and sandy mudstone (Pl.193). The mudstone consists of fine grained chlorite and muscovite with subsidiary quartz. Sandy mudstone, more common than silty mudstone, contains laminae, lenses, and thin, lenticular beds of ripple fine and very fine sandstone. The proportion of sand, and of lenses and laminae of sand is highly variable. At Trollfjord, a few large calcareous concretions were observed beneath the festoon cross-bedded sandstone in facies C (fig. 83).

Facies B2

This facies consists of thin and medium bedded sandstones almost always sharp and planar based, and occasionally highly erosive and lenticular. Tops of beds are usually sharp and are rippled at Trollfjord. The beds are invariably intercalated with mudstone of facies B1 (Pl.194). Individual beds range in grain size from fine to very coarse sand. At Trollfjord the beds contain numerous shale flakes, not common elsewhere. Coarse sandstone is usually massive, while fine and medium sandstone is parallel-laminated, or cross-bedded in sets from 5-25 cm thick, usually of trough-fill variety. Grading is rare.

Facies B2 always contains intercalated mudstone; as the proportion of sandstone decreases, the facies grades into facies B1. Where facies B1 is particularly sandy confusion between the two facies is possible, but the lenticular, rippled sandstones, with wavy bases and top contrasts with the more even bedding in facies B2 sandstones.

Coarse Sand Wave Beds

One variety of facies B2 is composed of very coarse sand and granules and exhibits a large-scale wave-like morphology which is believed to be a fossil antidune structure. Examples of the structure are in figure 83.

At Lavvonjargga the structures are about 80 cm in length and about 14 cm high (Pl.195). The lower part of the bed has cross-stratification dipping to the NE, while the crests have cross-stratification dipping to the SW (Pl.196). The structures are rounded and symmetrical, and one crest has a small scour with the sand from the scour displaced a few centimetres to the SW (Pl.196).

Similar structures were seen at Nesvatn where there is a row of disconnected lenses, internally cross-bedded, laminae dipping to the north. At Trollfjord three bedding plane surfaces with the structure were observed. The middle occurrence (fig. 83) consists of waves with slightly sinuous, relatively sharp crests. The bed is only weakly laminated. The highest occurrence is on top of a thick bed in facies C which nearby is festoon trough cross-bedded internally. These structures are relatively straight crested.

Facies B: Interpretation

Facies B1 formed by the fallout of mud and fine sand from suspension in a quiet, and occasionally agitated subaqueous floodplain environment. Rounding and reactivation of ripples (similar to internal erosion surfaces described by Collinson (1970)) indicate variable conditions, and wave activity, and the absence of raindrop imprints and mud cracks suggest subaqueous conditions.

Facies B2 represents the incursion of coarser sandstones into the quiet environment. Strong currents are indicated by the sharp or erosive bases, shale flakes, and coarse grain size with the sedimentary structures suggesting flow in the upper part of the lower flow regime, and upper flow regime. Rare grading testifies to gradual waning of the current, but the sharp top of most beds suggests that this was relatively rapid. Rippled tops of some beds at Trollfjord suggests reworking by currents and waves.

An antidune origin for the sand waves in facies B seems likely as: 1) the coarse grain size is consistent with emplacement with strong currents, likely to occur during flooding; 2) antidunes have a rounded and symmetrical form similar to the structures observed (Middleton, 1965), and similar to other structures considered to be antidune deposits (Hand, 1969), and 3) antidunes move in both upcurrent and downcurrent directions, and thus cross-bedding in two directions may form from one unidirectional current.

Of the alternative mechanisms that can be invoked to explain the sand wave beds, wave activity is unlikely because in a floodplain environment coarse grain sizes are associated with strong, spasmodic currents, rather than waves which tend to rework fine sand and silt. The migration of dunes in the upper part of the lower flow regime is also unlikely as a symmetrical profile would be attributed to rounding of the dune during a period of erosion. Such erosion has been described from the channel of the Tana River (Collinson, 1970), where the internal cross-stratification of the dune is truncated by a much more gently inclined erosion surface (reactivation surface). In the example at Trollfjord, the upper surface is concordant with the internal cross-stratification.

A relationship between wavelength and velocity for antidunes (Middleton, 1965) where $u^2 = gL/2\pi$, u = velocity, g = gravitational constant, and L = antidune wavelength, indicates that current velocity was about 112 cm/sec. A relationship between wavelength and depth (d), $L = 2\pi d$ (Allen, 1970, p.82) indicates that the maximum water depth was about 16 cm.

These antidune structures are unusual in that the internal structure is clearly preserved, possibly due to the coarse grain size. The sand wave form in facies C at Trollfjord may have formed by wave activity.

Facies B1 is attributed to the normal quiet conditions of the submerged floodplain, while facies B2 appears to be crevasse deposits. During floods, strong currents may carry coarse sediment, normally channel bed load onto the floodplain, where it is rapidly deposited. The alternating sand and mud is typical of floodplain deposits (Allen, 1965; Schumm and Lichty, 1963), the mud deposited from quiet water during the waning of the flood (e.g. Schumm and Lichty, 1963).

Variation in the ratio between B1 and B2 in a section may depend on several parameters, which could not be determined in the present study because of the absence of laterally continuous exposures. These might include proximity to channel, floodplain topography, and local channel morphology.

The rapid switch from channel to floodplain conditions suggests that the river moved laterally by avulsion, rather than by slow migration, supporting this aspect of Moody-Stuart's (1966) rather than Allen's (1965) model of straight channel streams.

8.4.5 Facies C and the Transition to Marine Conditions

Description

Facies C includes both sandstones and shales. Towards the top of most sections, the rippled, sandy mudstones of facies B1 grades upwards into a finely laminated grey mudstone, and then into laminated purple mudstone. These finely laminated mudstones are considered as part of facies C. In addition sandstones distinct from the medium grained, poorly sorted, micaceous sandstones in the lower part of the section appear in most sections intercalated with the finely laminated mudstone. The sandstone, also part of facies C, is relatively well sorted and well rounded, fine to medium grained, occasionally coarse grained, generally in units 2-6 m thick with sharp planar bases and tops. Internally they are parallel-laminated, festoon trough cross-bedded, low-angle cross-bedded, horizontally bedded or apparently massive (fig. 83).

In addition to facies C, two features of facies A and B which occur 20-30 m from the base of the upper submember are small-scale coarsening upwards sequences and small scours in facies B1. Small-scale coarsening upwards sequences 3-5 m thick were observed in the Leirpollen section (fig. 83, about 20 m from the base) and south of Melkelven (Pl.197). Rippled and parallel-laminated silty mudstone at the base of the sequences passes up into thin medium bedded, medium grained sandstones as the frequency, thickness and grain size of the sandstones increase gradually upwards at the expense of the mudstones. These are overlain sharply or erosively by medium to thick bedded, medium to coarse grained, massive and cross-bedded sandstones similar in appearance to those of facies A.

In its upper part, facies B1 may contain small scours up to about 1 m deep with smooth gently sloping, convex down bases, cut into and filled with the same rippled and parallel-laminated sandy and silty mudstones. These were particularly well developed south of Melkelven. The laminated fill either drapes up, concordantly, over the margins of the scours, or is cut out against the margins.

Interpretation

Facies C sandstones appear to have undergone a stage of winnowing of the mud fraction not found in the fluvial channel sandstones of facies A, and which may have been due to marine processes. Marine transport and reworking of fluvially derived sands may result in the formation of spits, beaches, bars and cheniers in a deltaic framework, either during the constructive (Morgan, 1970) or destructive phases (Scruton, 1960; Curtis, 1970). Mississippi delta chenier deposits are thin and the stratification is gently dipping, (Hoyt, 1969). Sand is mainly fine, well sorted and rests sharply on the underlying sandy and silty clays (Byrne et al., 1959). The shoestring geometry would impart a lenticular appearance in a transverse cross-section. These aspects fit well with the facies C sandstones, but cheniers in the present day are associated with a steady relative sea-level, while it appears that there was relative rise in sea-level while the upper part of the upper submember was deposited. Alternatively, spits and beaches tend to form where marine currents cause progradation adjacent to, or a considerable distance from distributaries. Such progradation into a quiet area such as an interdistributary bay or an offshore area forms a gradually coarsening upwards sequence (Fisk, 1955; Oomkens, 1970). The absence of such sequences beneath facies C sandstones suggests that the beach and spit origins are not correct. Insufficient exposure does not permit the detailed interpretation of the facies C sandstones; but the importance of marine influence in their formation is likely.

The coarsening upwards sequences appear to represent a transition from quiet water sedimentation facies B1 to conditions of strong currents in channels (fig. 83, Leirpollen). Such sequences are virtually unknown in ancient deposits attributed to a fluvial origin, or in present day fluvial sediments. On the other hand, such sequences have been described from the Rhone delta where they have been termed fluviolacustrine offlap sequences (Oomkens, 1970) and may have formed by the progradation of a minor distributary channel and minor mouth bar into interdistributary bay (Coleman, et al., 1964). According to Oomkens (1970, p.205) small-scale coarsening upwards sequences can form where there is a density difference between the incoming water and the basin water so that a jet can form. However in the absence of the density difference, rapid mixing will take place with little transport, causing the blocking up of the crevasse channel. This appears to explain the absence of such sequences in fluvial deposits. The sequences thus suggest that facies B towards the top of the member represents brackish interdistributary bay, rather than fresh-water floodplain.

Scours cut into and filled with facies B1 suggest periods of high energy, erosive currents without the influx of correspondingly coarse sediment. Similar scours floored with a graded bed of coarse, poorly sorted sandstone occur opposite Areholmen (Banks, 1971). Banks interpreted these sandstone beds as river generated, forming when the river was in flood (presumably some of the channels were cut at the same time), while the siltstone (facies B1) filling channels was interpreted as intertidal channel deposits accumulating at the margin of an estuary. The present author observed no direct evidence for subaerial exposure, and thus tends to favour a subtidal origin for this facies.

8.5 Discussion

8.5.1 Significance of the Facies Sequence A->B->C.

The facies sequence in most sections of the upper submember shows a broad pattern of fluvial channel sandstones at the base followed by floodplain-interdistributary sandstones and mudstones, followed by coastal marine mudstone and sandstone. Facies A is restricted to the lower 20 m of the submember, and in most sections only one horizon of channel facies occur, at the most two. The persistent thickness of about 40 m over the whole area, of the upper submember gives it sheet-like geometry, also noted by Reading and Walker (1966, p.197). This is a distinct contrast with the lower submember.

These considerations suggest that the upper submember formed as a result of a rising sea-level. The uniform thickness of the fluvial channel deposits suggests that aggradation occurred as a result of a gradual, persistent rise in base level. Had the preservation of the upper submember been due to regional subsidence we might have expected an irregular (perhaps wedge-shaped) thickness distribution, and repeated fluvial cycles where the thickness was appreciable. The origin proposed above necessitates a diachronous boundary between the Lillevatn and Innerelv Members, as was suggested by Reading and Walker (1966, p.208).

If fluvial deposition depended on a rise in base level, then what were the conditions of grade before the onset of the rise in base level? As the fluvial system appears to be related to the progradation of an integrated delta slope and platform, it seems unlikely that entrenchment of streams took place; this is supported by the absence of deep scours into the lower submember. Also, it would seem that all sediment was in transit, that permanent deposition was not occurring, and that residence times of individual particles was relatively short. Thus during progradation permanent deposition was mainly in the delta foresets, while during the transgression, fluvial and delta top deposition prevailed.

8.5.2 Palaeocurrent Evidence

Where numerous palaeocurrent directions could be recorded (fig. 83) it was observed that there was a great variability in currents from place to place in both facies A and B. Broadly, currents went in all directions except due east. This is not typical of many ancient fluvial channel deposits, nor is it a feature of models which have been proposed, either meandering or straight (Moody-Stuart, 1966; Allen, 1966). Sources of variance (Allen, 1966) are attributed, in a straight channel, to the meandering thalweg, variations in the orientation of dunes caused by the helicoidal flow, and a reverse eddy in the lee of the side bars. The latter part of Allen's explanation does not apply to the Lillevatn Member if the author's model is correct (fig. 87). Sufficient data were not collected to allow speculation on whether the variability has any hydrodynamic significance. However, the transgressive nature of the sequence may have been an important factor. In contrast to channel models, floodplain deposits may be expected to show a high variance (Allen, 1966).

The scarcity of currents to the east suggests that the generalized flow was from that direction, as was the source of the sediment.

8.5.3 Summary

In the south of the area prodelta deposition followed with rapid transition, glacial marine sedimentation, with about 40 m of prodelta sediments preserved (thick lower submember). To the north prodelta deposits are thin, 3-8 m, and they appear to have formed in relatively shallow water, following a period of reworking of the tillite. Both the thin and thick lower submembers are overlain erosively by the upper submember. The latter contains three groups of facies: at the base fluvial sandstones believed to have been deposited in straight channels (facies A) pass abruptly upwards into mudstones with intercalated sandstones which are interpreted as floodplain-interdistributary bay sediments (facies B). These are succeeded by mudstone (facies B1 and C) with relatively clean sandstones (facies C) which may have formed in coastal environments. Small-scale coarsening upwards sequences in the middle of the upper submember indicate brackish or saline conditions and suggests that the sediments may be partially of interdistributary bay origin. The sheet-like geometry of the upper submember, the restriction of fluvial channel deposits to the base of the upper submember, and the passage up into marine sediments suggest that the upper submember formed as a response to a rise in sea-level. Palaeocurrents are highly variable, but suggest that the sediment was brought into the area from the east.

Key to Figure 83.

Grain size

Sh = mudstone

Slt = siltstone

Ss = sandstone

Facies

A = channel

B = floodplain and interdistributary bay

C = coastal plain and marine influence

Contacts

sharp

erosive

gradational

Sedimentary Structures

Bedding

Cross-bedding

Parallel lamination

Cross-lamination

Mudstone laminae or partings

Massive

Sand wave form

Palaeocurrents

channel margin

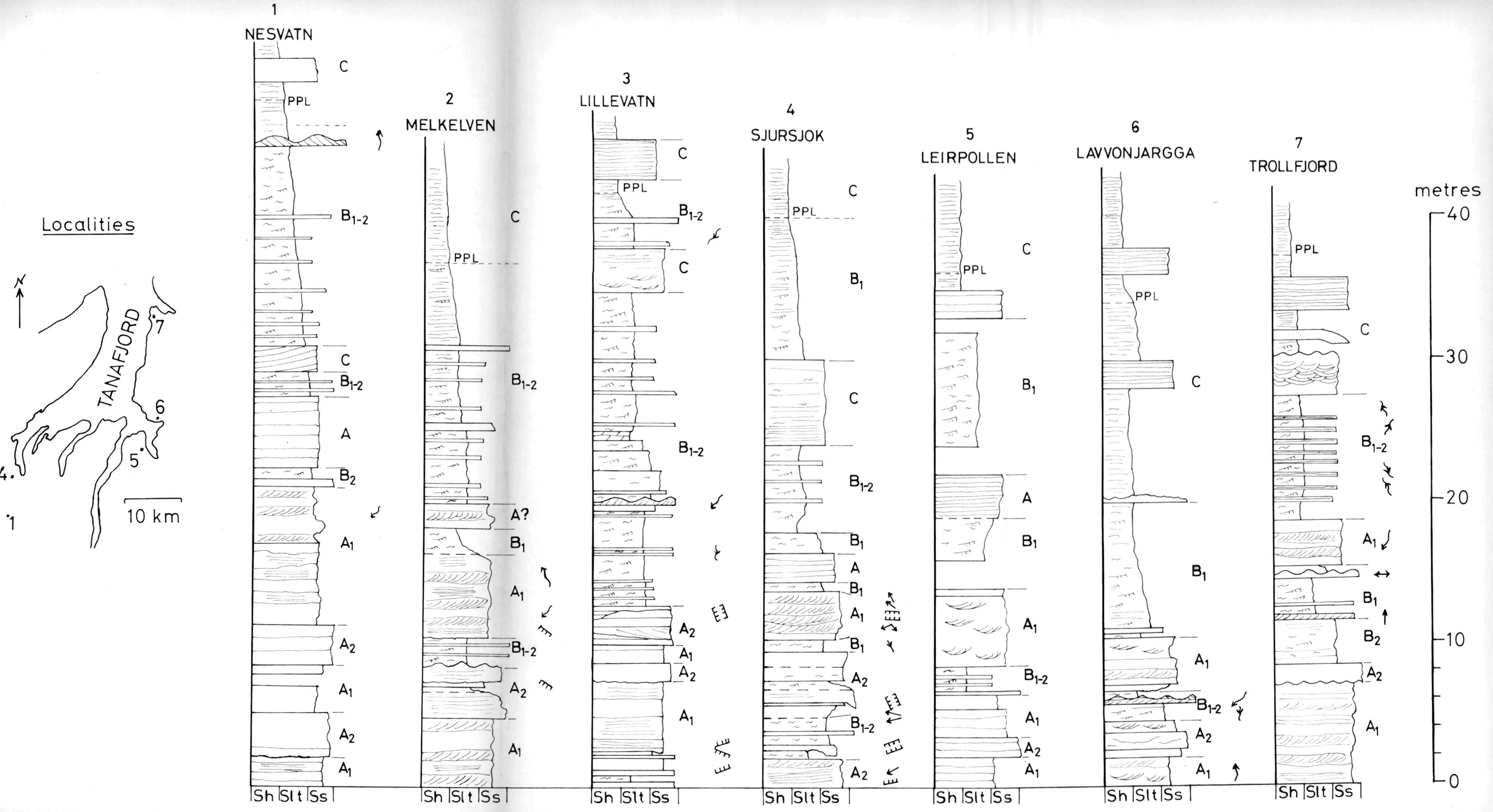
dip of cross-bed

asymmetrical ripple

ripple or dune crest

wavy line indicates approximate direction

North is towards the top of the page.



Localities
TANAFJORD
10 km
1
NESVATN
2
MELKELVEN
3
LILLEVATN
4
SJURSJOK
5
LEIRPOLLEN
6
LAVVONJARGGA
7
TROLLFJORD
PPL
C
A
A?
B_{1-2}
B_1
B_2
A_1
A_2
Sh
Slt
Ss
metres
0
10
20
30
40