SIGNIFICANCE OF SWALEY CROSS-STRATIFICATION IN THE CARSTONE FORMATION, HUNSTANTON, NORFOLK

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ABSTRACT

Cross-bedding in the Lower Cretaceous Carstone Formation at Hunstanton cliffs in Norfolk was first reported almost a hundred years ago, but details on the nature and significance of the cross-bedding have been largely ignored. To redress this we made detailed field observations and measurements. Fourteen cross-bedded sets were suitable for accurate measurements of set thicknesses, visible set length and foreset orientation. Eleven cross-bedded sets had NE trending foreset dip directions, while three sets had S trending foreset dip directions. Circular statistics on the directional data gave a significant NE mean dip direction of 55°. Cross-bedding geometry is broadly of 'trough-type' but specifically 'swaley', characterised by concave-upward shallow scours between 0.5-2.0 m wide and a few centimetres deep. Swaley cross-stratification is thought to form below fair-weather wave base, but above storm wave base; our calculations using published physical equations suggest the Carstone bedforms were generated on a storm dominated shoreface in 30 to 40 m water depth. Northward palaeocurrent directions are strikingly different to the predominantly southward palaeocurrents recorded in older Lower Greensand deposits of southern England. A storm surge relaxation current from the land deflected right by Coriolis forces could have resulted in an alongshore NE trending flow forming the Carstone cross-bedding. This interpretation allows for predominantly southward ocean currents interrupted by episodic storms surges and resulting relaxation currents.

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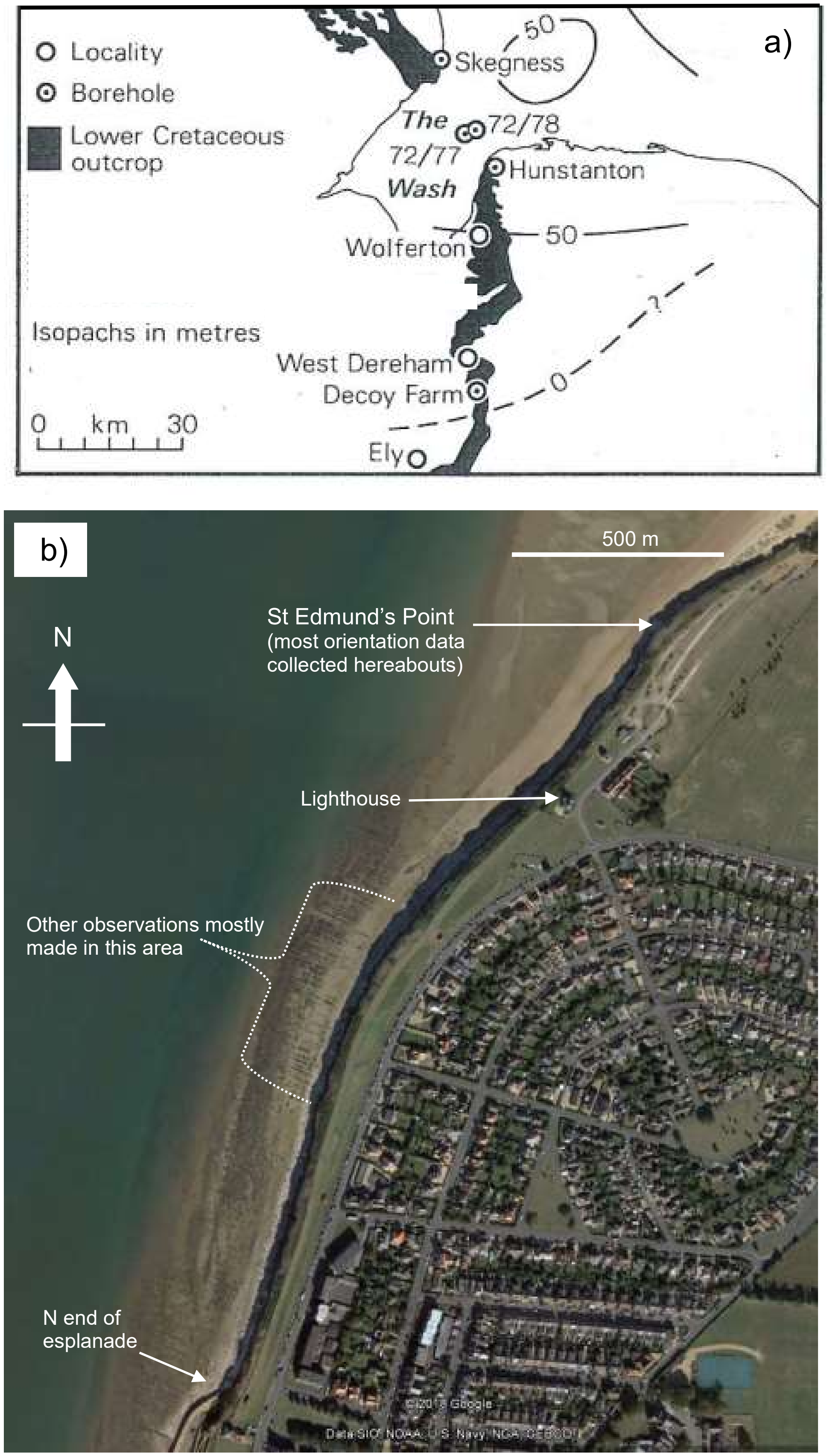
INTRODUCTION

Sedimentary structures in the Lower Greensand Group of southern England have been described in both sedimentological and sequence stratigraphic contexts (Schwarzacher, 1953; Narayan, 1971; Dike, 1972; Ruffell & Wach, 1991; 1998a; 1998b; Eyers, 1995; Woods *et al*., 2001). Conversely, the significance of cross-bedding in the near-time equivalent Lower Cretaceous Carstone Formation of west Norfolk has been either missed (Schwarzacher, 1953), or mentioned only briefly (Versey & Carter, 1926; Andrews, 1983; Gallois, 1994; Thomas, 1998). The research described here thus sets out to describe and interpret the nature of cross-bedding in the Carstone Formation to help refine palaeoenvironmental interpretations and to understand how reconstructed palaeocurrent directions might fit into the wider context of Lower Greensand Group sedimentology.

The study site was the cliff section at Hunstanton (Fig. 1a) between the Esplanade [NGR 6723 4124] and St Edmund's Point [NGR 6786 4239; Fig. 1b]. Here the upper 10 m of the Middle to Upper Albian Carstone Formation (Owen, 1995; Fig. 2) is well exposed both laterally (~1400 m) and vertically, allowing close examination of bedding geometry and sedimentary structures. The structural dip of bedding in the cliff sequence is to NE, such that although the cliffs are up to 18 m high, at their northern end the top of the formation and its junction with the overlying Hunstanton Formation (Fig. 2) is brought to eye level where it can be examined in detail.

Fig. 1. (Opposite page). a) Map showing position of Hunstanton in relation to outcrop of Lower Cretaceous rocks in Norfolk and south Lincolnshire with isopachs of pre-Albian sediments (modified from Rawson, 1992). b) Google Earth image of north Hunstanton showing the main areas where observations and measurements were made in the cliffs. Data SIO, NOAA, U.S. Navy, NGA GEBCO, image © 2018 TerraMetrics.


a)
Locality
Borehole
Lower Cretaceous outcrop
Isopachs in metres
0 km 30
50
Skegness
The
72/78
72/77
Wash
Hunstanton
Wolferton
50
?
West Dereham
Decoy Farm
0
Ely
b)
500 m
N
St Edmund's Point
(most orientation data collected hereabouts)
Lighthouse
Other observations mostly made in this area
N end of esplanade

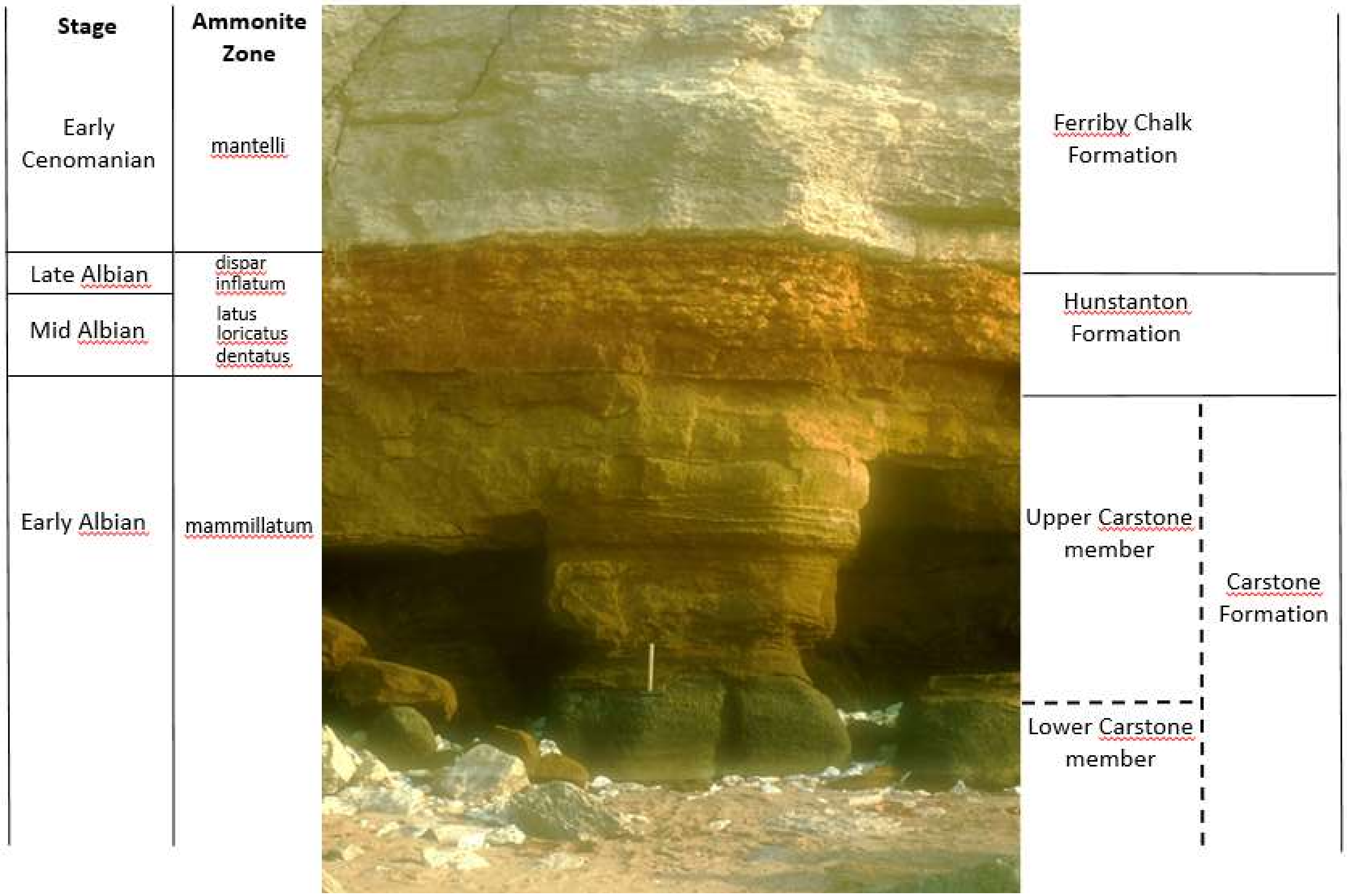


Fig. 2. Local stratigraphy of the Hunstanton cliff section at St Edmund's Point [NGR 6786 4239; Fig. 1b]. The hammer (handle 35 cm long) rests on the boundary between the Lower and Upper Carstone members (informal names after Owen (1995). The top of the Lower Carstone member seen here is a dark brown, iron-oxide cemented, cross-bedded conglomerate, the main unit of study in this paper. Stage and ammonite zones after Hopson *et al*. (2008).

The Carstone rests unconformably on older Cretaceous strata when traced southwards in north Norfolk. Further south still it rests on Jurassic strata, and eventually on Silurian – Devonian basement rocks of the London-Brabant High, proved by boreholes around Diss and Bury St Edmunds (Wilkinson, 2006). Around Ely (Fig. 1a), Gallois (1988) has shown that the Carstone Formation has an erosive contact with the Aptian Lower Greensand (Woburn Sands Formation), and is thus stratigraphically younger.

The Lower Greensand Group of southern England is characterised by coastal, lagoonal and shallow marine shelf lithofacies (Fig. 4), dominated by sandstones with subordinate silts and clays in some intervals (Wilkinson, 2006). These sediments mark a major Lower Cretaceous transgression that ended a 40 million year period of predominantly non-marine deposition represented by the Wealden facies of southern England (Ruffell & Wach, 1991). The Lower Greensand Group is best developed (250 m thick) on the Isle of Wight (Anderton *et al*., 1979) serving as a standard of reference for the Channel (Vectian) Basin and the type locality for the standard ammonite zones of the English Aptian. Coarser grained facies in the Lower Greensand Group are characterised by southward trending large-scale cross-bedding. Between Sandy and Leighton Buzzard in Bedfordshire, cross-bedded sets in the Woburn Sand Formation (Fig. 3) are between 1 to 10 m thick and indicate currents flowing predominantly from the north (Schwarzacher 1953; Shephard-Thorn et al., 1994; Bristow, 1995). The cross-bedding is interpreted as tidally influenced sand-wave complexes that migrated in a south-easterly direction through the 'Bedfordshire Straits' (Rawson, 1992; Fig. 4). Similar large-scale cross-bedding is present in other parts of the Group (e.g. the Hythe, Sandgate, Ferruginous Sands and Sandrock Formations; Narayan, 1971; Dike, 1972; Ruffell & Wach, 1998b; Fig. 3) in southern England. Large-scale (2-4 m thick) cross-bedding is weakly developed in parts of the Monk's Bay Formation (Dike, 1972; Hopson *et al*., 2011; stratigraphic equivalent of the Norfolk Carstone Formation; Fig. 3), again interpreted as tidally influenced sand-wave facies (Dike, 1972) although foreset dip directions are not recorded.

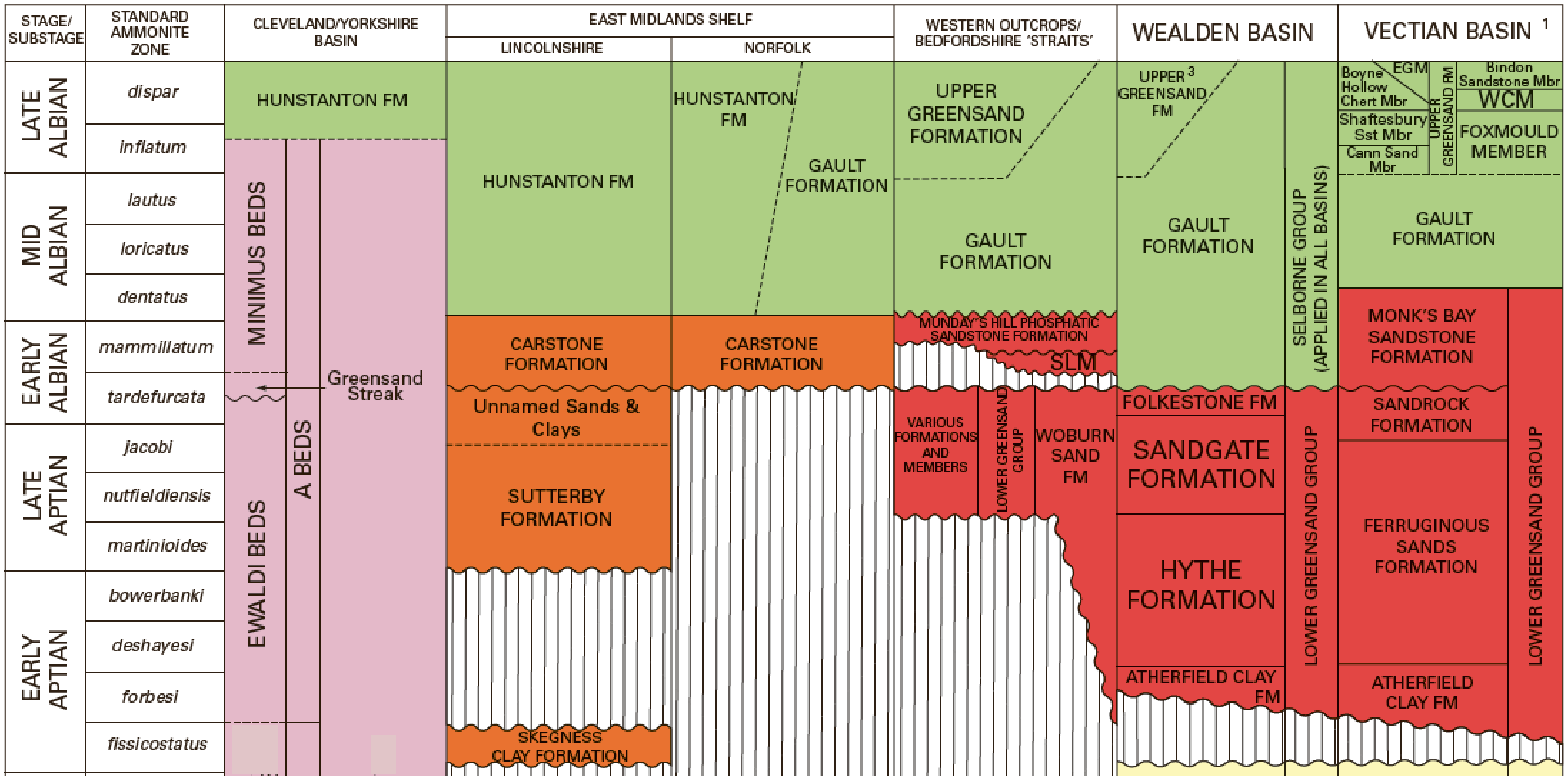


Fig. 3. English Aptian to Albian Stratigraphy adapted from Hopson et al. (2008) BGS Research Report RR/08/03, with the permission of the British Geological Survey. [1]The Vectian Basin has also been called the Channel Basin by Ruffell & Wach (1998b). The Monk's Bay Sandstone Formation is the new name for the Carstone on the Isle of Wight.

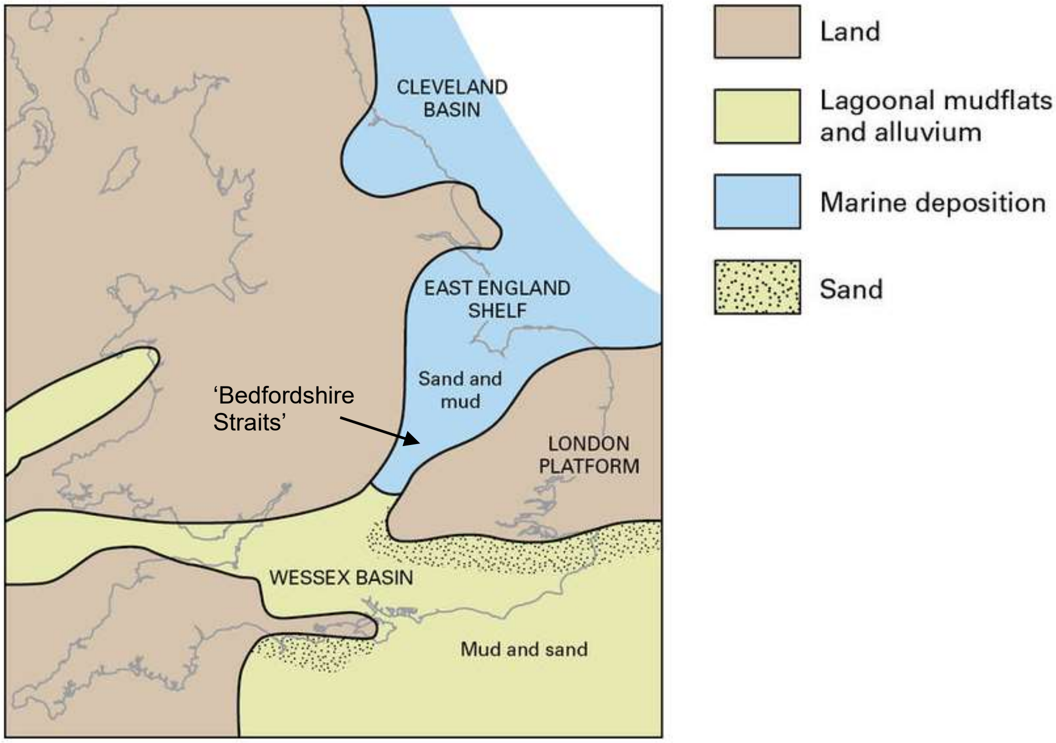


Fig. 4. Aptian-Albian palaeogeography of England based on Rawson (1992). As the Aptian-Albian transgression progressed the southern England lagoons and mudflats become tidal estuarine and clastic shallow shelf environments (see Fig. 12). Base image P785825.jpg courtesy of the British Geological Survey.

THE CARSTONE FORMATION

The Carstone Formation is 18.9 m thick in the Hunstanton Borehole, while at the cliff the exposed thickness is ~10 m (Gallois, 1994). The formation thins to the south (between 1 to 7 m) and north-west, beneath the Wash where it is ~6 m thick (Gallois, 1994).

At Hunstanton Cliff the Carstone Formation has been divided into two informal members (Owen, 1995; Wilkinson, 2006; Fig. 2). The Upper Carstone member (~2.60 m thick) is a medium grained, orange-brown, ferruginous, in part oolitic (chamosite), friable sandstone which contains small phosphatic nodules. While body fossils are very rare in the Carstone (Gallois, 1994) these phosphate nodules have yielded two ammonites, both specimens of *Hoplites dentatus* (J.

Sowerby) (Owen, 1995). The ammonites indicate that the top of the Carstone is of Middle Albian, *Hoplites spathi* Subzone age (Owen, 1995). These upper beds also contain evidence of bioturbation, fresh exposures displaying *Arenicolites* and *Skolithos* type burrows (Gallois, 1994), infilled with phosphatised sand. The Lower Carstone member, seen to about 7.0 m (base below modern beach sand), is a dark brown, friable pebble conglomerate, cross-bedded, with coarse to very coarse, poorly sorted, sandy matrix cemented by iron-oxide (Fig. 2).

The Carstone Formation sediments fine upward, passing conformably into the overlying red chalks of the Hunstanton Formation (Fig. 2) and both have long been regarded as current-swept shallow marine deposits (Owen, 1995) formed under transgressive conditions. The change to chalky lithologies (Hunstanton Formation) is accompanied by the presence of marine body fossils including ammonites, belemnites, nautiloids, bivalves and brachiopods (Gallois, 1994; Owen, 1995; Mitchell, 1995; Underwood & Mitchell, 1999; Price & Harwood, 2012; Andrews *et al*., 2014). The Hunstanton Formation is in turn overlain by the Cenomanian Ferriby Chalk Formation (Hopson, 2005; Fig. 2).

Sedimentary structures including cross-bedding are probably present throughout the Carstone Formation, but the exposed cliff section is deeply weathered such that primary structures are mainly obscured by secondary, limonitic, weathering products (Gallois, 1994).

CROSS STRATIFICATION

Cross-stratification is a sedimentary structure found in marine, fluvial and aeolian siltstones, sandstones or conglomerates, formed by downstream migration of bedforms. Sediment is deposited on the upstream (stoss) side of a ripple or dune and after reaching the angle of repose, the grains avalanche down the downstream (lee) side. Repetition of this process creates a distinctive internal layering in the bedform, which can be preserved, usually in two dimensions in sedimentary rocks. The individual lee face layers are known as foresets; these depositional features dip in the direction of sediment transport and thus preserve the direction of palaeocurrent (Fig. 5).

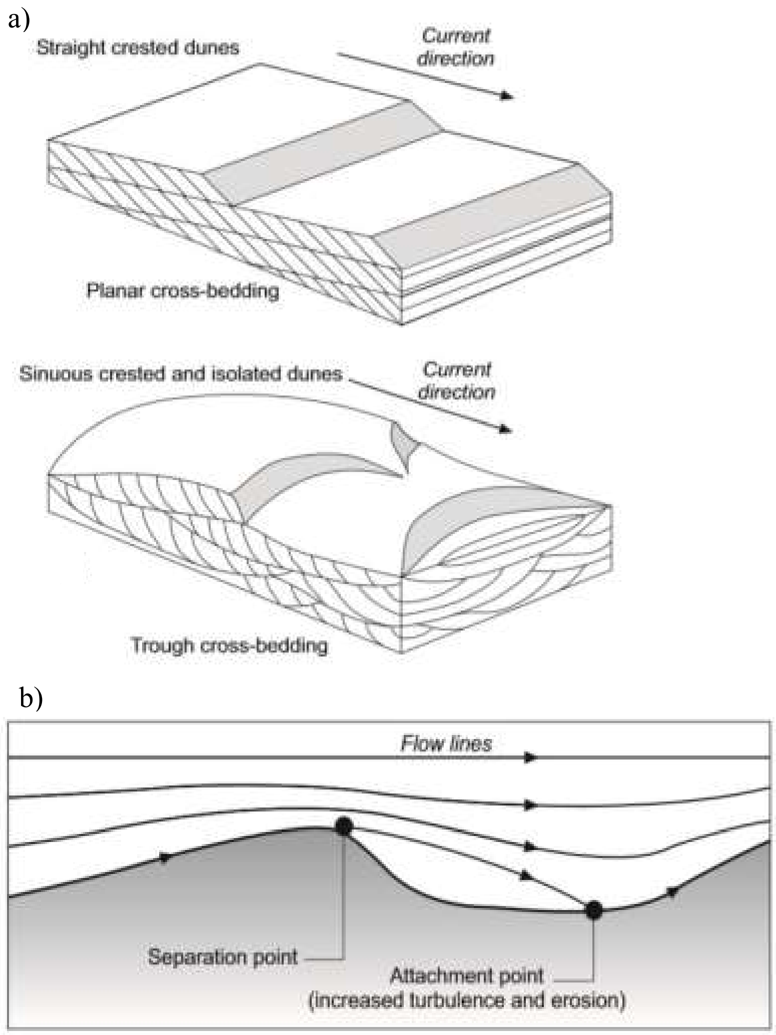


Fig. 5. a) Migrating dunes forming planar or trough cross-bedding. b) Fluid flow over a bedform showing imaginary streamlines (after Nichols, 2009).

Ripples and dunes migrate by the removal of sediment from the upstream side followed by deposition on the downstream side. Sediment supplied by scour on the upstream side moves up the stoss slope of the next dune creating a train of dune troughs and crests advancing downstream (Nichols, 2009). If there is net addition of sediment, deposition on the lee side will be greater than net removal from the stoss side. The dune will then grow and the depth of stoss side scour is reduced, preserving the cross stratification created by previously migrating dunes (Leeder, 2011; Fig. 5a). The geometry of bedforms is a function of hydrodynamic conditions and physical parameters of the sediment such as grain size, sorting, shape and density (Baas, 1994). The thickness of cross-stratified sets preserved in rocks ranges from a few centimetres to several metres depending on sediment supply, hydrodynamic conditions and post-depositional erosion and deformation.

The style of cross-stratification depends on the predominant conditions during deposition (Fig. 5a). Tabular or planar-cross stratification describes low sinuosity bedform crests, the cross-beds bounded by horizontal surfaces. The bottom of these cross-beds is often close to horizontal because of the absence of scouring in the trough (Nichols, 2009). Stronger flow may erode a scour pit at the reattachment point (Fig. 5b) creating an undulating erosion surface, the base of a 'trough' or 'swale'. Under these conditions the dune crest is highly sinuous and trough cross-bedding is formed by the migration of these sinuous crested dunes; such cross-bedding typically has an undulating lower boundary with asymptotic foreset contacts (Fig. 5a; Nichols, 2009).

Because of the variance in the type and size of cross-stratified sets the terminology used to describe them is diverse (see McKee & Weir, 1953). In this paper, the term "cross-bedding" refers to cross-stratified sets with a thickness of 5 cm or more. "Large-scale cross-bedding" refers to cross-stratified sets with a thickness over 1 m. "Cross-stratification" is used as a generic term to refer to these sedimentary structures without any size attribution.

METHODS

Lithologies, bed thicknesses, nature of bedding contacts, sediment grain size and sorting, presence of sedimentary structures, presence of trace fossils and evidence for bioturbation were noted and photographed in the field. Dip and strike of both structural bedding and cross-bedding were measured using a compass-clinometer. All observations were located on the Ordnance Survey grid using a handheld GPS.

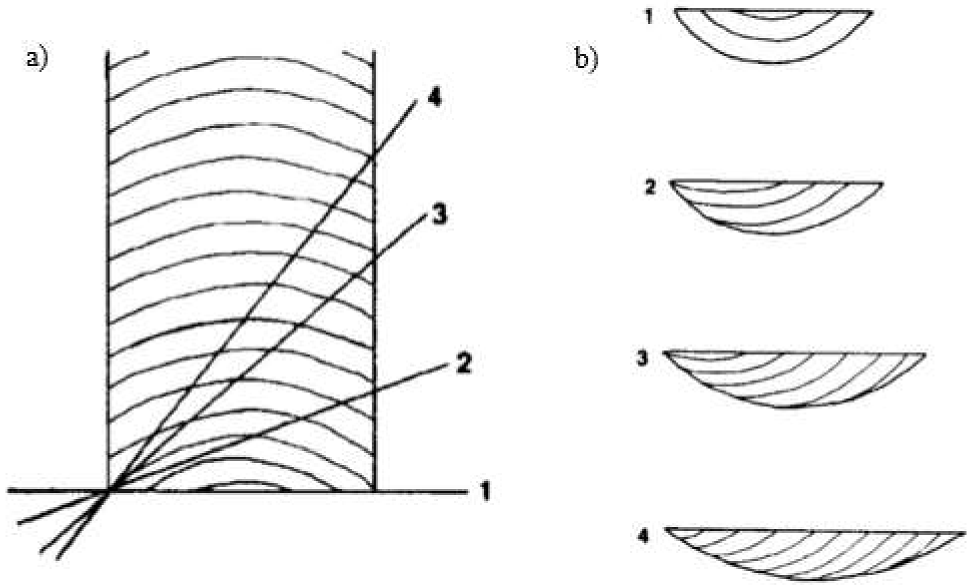


Fig. 6. a) Plan view of trough-cross-beds with different vertical cross sections. b) cross sectional view. Note how the apparent width of the trough and the number of truncated foresets and their dip angle increases as the cut approaches the longitudinal cut (after Decelles *et al*., 1983).

Field directional data from sedimentary structures is typically a two dimensional representation of a three dimensional structure. This is important in the case of cross-stratification because, unless the plane of the exposure is parallel to the direction of maximum dip, the measured dip angle will be lower (apparent dip) than the true dip (Fig. 6). While the uneven and weathered surface of the Carstone outcrop often made cross-bedding difficult to identify, it did allow examination from different angles, yielding quite accurate information on the three dimensional geometry

Thickness of cross-bedded sets and the length over which they are visible were recorded, while orientation of the cross-bedding was determined using a compass-clinometer. Fourteen cross-bedded sets were examined in detail where accurate measurements could be undertaken typically around modern beach level. For each cross-bedded set, two apparent dips were recorded with an accuracy of +/-

0.5°. These were later combined to ascertain the true dip and strike of the foresets. Thus, 28 directional measurements reveal the true orientation of 14 cross-bedded sets. The two apparent dip angles and directions were plotted on a stereographic net as point poles. The plane that bisects both points represents the true dip and strike of the foresets. All measurements were corrected for magnetic declination and for tectonic dip using stereographic nets. The latter correction had trivial impact on the directional data, changing the strike direction of the cross bedding by <1° as the structural dip at Hunstanton is ~4° (see also Lindholm, 1987).

Thirty-nine grain azimuth and plunge values within the cross-bedded sets were recorded using a compass-clinometer. The high degree of grain sphericity made longest axis identification difficult and measuring dip angle and direction of even elongated grains was challenging due to their small size. These data were plotted on a rose diagram to allow comparison with foreset orientation.

Circular statistics were used to determine the mean dip direction of the foresets and to test the significance of the mean. Directional statistics are sensitive to small sample sizes and it is important to test the directional mean for significance (Zar, 1981). The same method was used to determine the mean orientation of the grains and to test its significance. Grain orientation in conjunction with the cross-bedding orientation should provide a reliable indicator of palaeocurrent direction (Decelles *et al*., 1983).

RESULTS

Observations on cross-bedding

Observations on cross-bedding were made in the upper beds of the Lower Carstone member (Fig. 2) and examples are shown in Figures 7 & 8. Alternation of darker and lighter laminae typically make the cross-bedding visible, with cross-bedded set thicknesses ranging between 9 and 16 cm (mean 14 cm). Individual foreset layers were typically ~5 mm thick. Of the 14 cross-bedded sets studied in detail, 11 had foreset dip directions trending toward NE while 3 had foreset dip directions trending toward S (Appendix 1). Circular statistical analysis of the directional data

Fig. 7. Swaley cross-stratification seen at outcrop near St Edmund's Point [TF 6786 4239] ~3 m below top of Carstone Fm. Notebook is 20 cm long. At this locality the apparent foreset dips appear planar (see text and Fig. 10). Palaeocurrent direction is right to left, which is broadly toward NE.

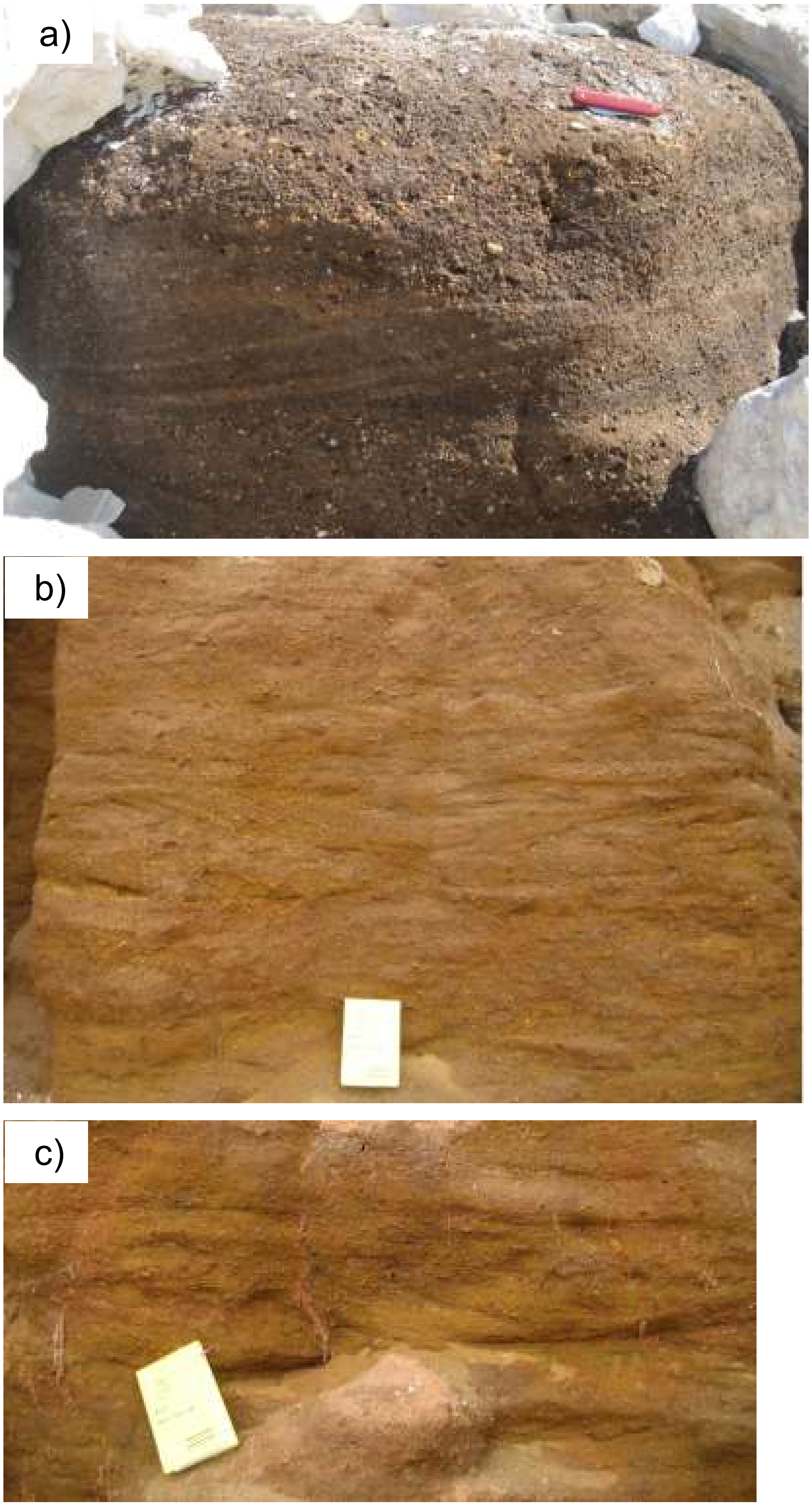


Fig. 8. Swaley cross-stratification seen at outcrop. Note the distinctive concave-upward shallow scours that delineate swales (see Fig. 10). Notebook is 20 cm long; penknife is 10 cm long. a) St Edmund's Point [TF 6786 4239] ~3 m below top of Carstone Fm; b & c) ~TF 6750 4180, ~8 m below top of Carstone Fm.

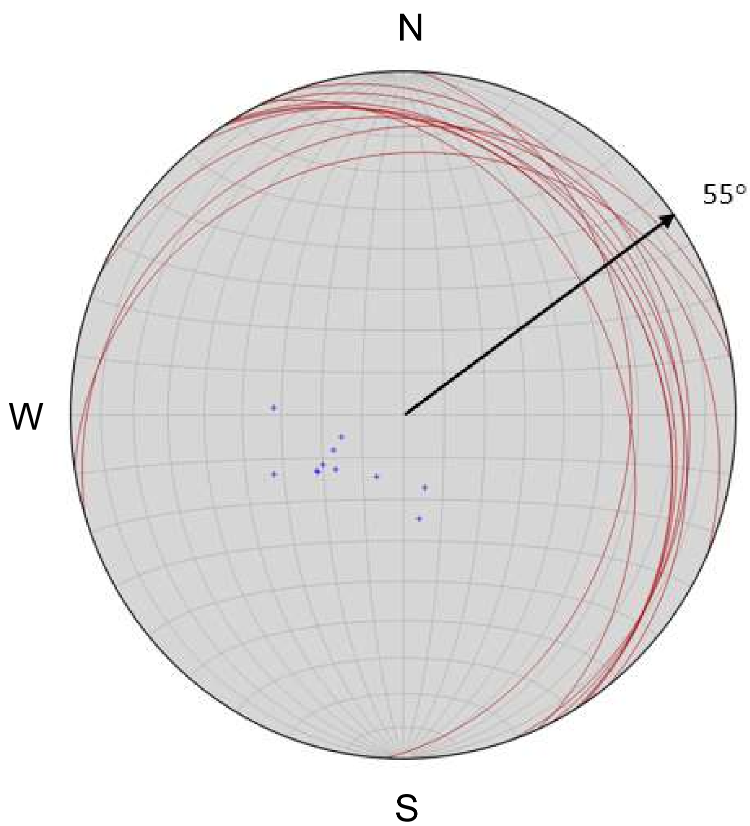


Fig. 9. Stereographic projection of eleven sets of N trending foreset dip directions showing vector mean of all data.

gave a statistically significant mean dip direction for the NE trending sets of 55° (Rayleigh z score = 9.08; 9.08>6.085 thus p<0.001) (Fig. 9). For the S trending sets the mean direction was 204° but the sample size was too small for significance testing. The dip angle of the cross bedding varied from 14° to 30° with an average dip angle of 23°.

The top and bottom of the cross-beds are bounded by erosion surfaces. The tops of foresets are thus truncated, meaning the total thickness of the cross-beds was greater than that preserved. Laterally, the truncation surfaces become conformable and the foresets here were asymptotic in nature, such that inclination along one foreset lamina decreased in the downcurrent direction, meeting the lower bounding surface at low angle.

Cross-bedding type

The observed cross-bedding geometry is broadly of 'trough-type' but specifically 'swaley' in nature, corroborating the findings of Thomas (1998). The term swaley cross-stratification (Leckie & Walker, 1982) characterises concave-upward shallow scours between 0.5-2.0 m wide and few centimetres deep (Fig. 10) and compare with outcrop structures in the Carstone (Figs 7 & 8), the thickening and thinning of laminae resulting in fan-like stratification with fluctuating dips.

Grain Orientation

The grain orientation data (Appendix 1) support the cross-bedding directional data. The average grain azimuth orientation is 27°, a significant value using a Rayleigh z test ($z=32$; $32 > 2.97$, $p<0.001$). The average grain dip angle was 15° with dip predominantly in the downflow direction (Fig. 11). While the grain orientation data show a similar trend to the cross-bedding directional data, their average direction is not as close to the plane of the cross-stratification as in some other studies. For example, Yagishita *et al*., (1992) analysed the grain fabric in swaley and hummocky cross-bedding and found that the preferred grain orientation was within 10° of the plane of the cross-laminae.

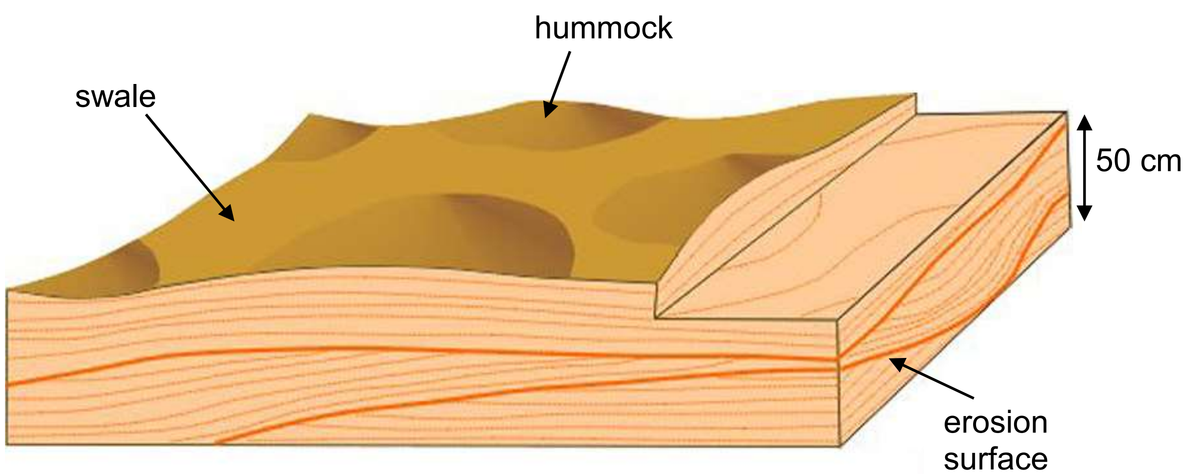


Fig. 10. Block diagram modifed from Nichols (2009) showing idealised hummocky and swaley cross-stratification, erosion surfaces and forsets. Note that in the frontal cross section the foresets can appear planar or even sub-horizontal (cf. Fig. 7).

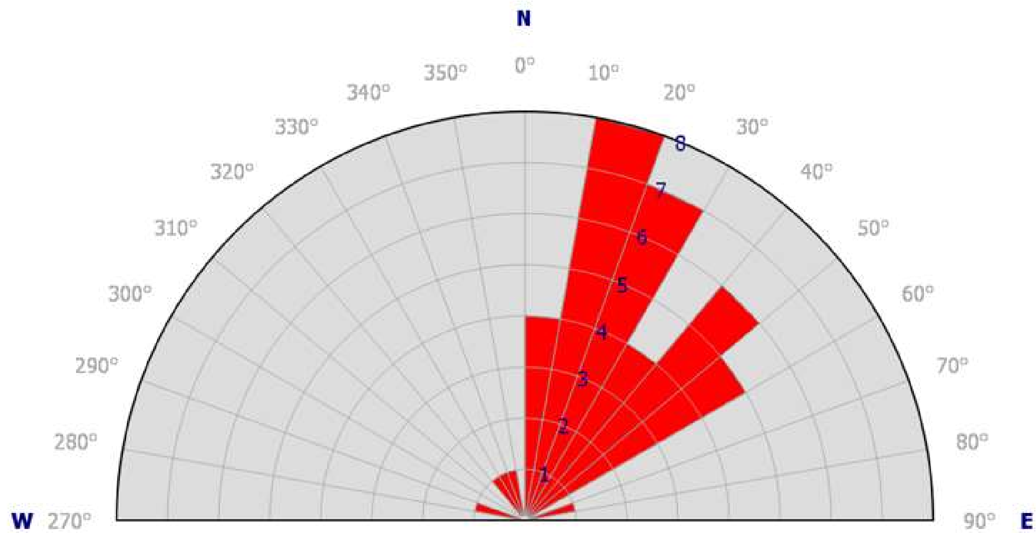


Fig. 11. Rose diagram (10° intervals) of grain dip direction. Number of samples (n) indicated by integers on the radius. Dataset n = 39.

DISCUSSION

Nature and stratigraphic distribution of cross-bedding

Hummocky and swaley cross-stratification are widely reported from the shallow-marine sedimentary record (Harms *et al.,* 1975; Leckie & Walker, 1982; Johnson & Baldwin, 1996) and there is consensus that both are diagnostic structures of storm-generated currents (Dott & Bourgeois, 1982; Walker, 1984; Duke *et al*., 1991; Morsilli & Pomar, 2012). These bedform types are genetically related, the main difference being the aggradation rate during their formation (Dumas & Arnott 2006). Under low aggradation rates (1 mm/min) swales are preferentially preserved and hummocks selectively eroded, while at higher aggradation rates, swales and hummocks are equally abundant. In the Carstone we observed only swaley cross bedding, suggesting that aggradation rates were relatively low. We note that swaley and hummocky cross stratification can appear planar depending on the plane of the cut (Figs 7 & 10). The characteristic swales are only evident if the plane of the cut is transverse to the trough axis. Most cuts in the Carstone are oblique to the trough axis, obscuring the characteristic concave upwards morphologies and giving the appearance of planar cross-bedding (Fig. 7) as recorded by Gallois (1994).

Cross-bedding was observed only in the coarser-grained, conglomerate facies (Lower Carstone member) and not in the sandstone facies (Upper Carstone member). This is slightly unusual as hummocky, swaley, or trough cross-bedding is more often seen in sandstones, especially in fine sand and silt facies (Harms *et al.,* 1975; Walker, 1984). The lack of preserved sedimentary structures in the Upper Carstone member (sandstone) does not necessarily mean that they were not formed, indeed the rather massive and uniform nature of the sandstones may simply imply intense bioturbation (Narayan, 1971; Woods *et al*., 2001; Wilkinson, 2006) that destroyed primary structures.

Swaley cross-stratification and palaeodepth estimates

Hummocky and swaley cross stratification is thought to form below fair-weather wave base, but above storm wave base (Harms *et al.,* 1975; Duke, 1985), typically in the transition between lower shoreface and offshore facies. Wave base describes the depth below which no significant motion occurs; sediment below this level is not transported by wave action on the surface. Hummocky dunes form immediately after the peak of a storm event due to fall out of suspended sediment and the moulding of the seabed by combined currents and long wave oscillations (Li & Amos, 1998). Laboratory experiments by Dumas and Arnott (2006) suggest that most rock record hummocky cross-stratification was generated by migrating bed forms under long wave periods (8-10 s) and oscillatory-dominant combined flow with a slight unidirectional component. In such conditions, relatively high oscillatory velocities (Uo > 50cm/s) and slow unidirectional currents dominate (Uu < 10cm/s). Swaley cross-stratification is believed to be formed under similar hydrodynamic conditions but at lower net aggradation rates, causing selective erosion of hummocks and preferential preservation of swales. This typically fixes the focus of swaley cross-stratification formation landward of hummocky cross-stratification, where higher net aggradation rates prevail (Dumas & Arnott, 2006). This is supported by the common observation of swaley cross-stratification stratigraphically above hummocky cross-stratification in shallowing upward shallow marine sequences (Leckie & Walker, 1982; Duke, 1985).

If we assume that swaley cross-stratification in the Carstone was formed under a wave period of ~9 seconds (Dumas & Arnott, 2006) the dispersion relationship can be used to calculate bedform wavelength based on wave period (Immenhauser, 2009).

For deep water (water depth > ½ wavelength) the relationship is:

Equation 1 $$L = \frac{gT^2}{2\pi}$$

where L refers to wavelength in metres, T to wave period in seconds, and g acceleration due to gravity (9.81 m s^{-2}). For a wave period of 9 seconds the resulting wavelength is 126 m. To calculate effective wave base, L is divided by 2 (Immenhauser, 2009), so for the Carstone, wave base is calculated to have been 63 m. Thus, waves with a wavelength of 126 m could only generate cross-bedding above 63 m water depth.

Dumas and Arnott (2006) determined that a minimum of 50 cm/s orbital velocity is required to generate hummocky cross-stratification. In order to determine the orbital velocity of water particles near the seabed, Curray (1960) used the following expression:

Equation 2 $$U = \frac{\pi H}{T} \times \frac{1}{\sinh \frac{2\pi d}{L}}$$

where U is the orbital velocity of water particles, H represents the height of the wave in metres, d is water depth in metres, T is wave period in seconds and L is wavelength in metres. Rearranging the equation to solve for depth, substituting an orbital velocity of 50 cm/s with a wave height of 4 m as suggested by Dumas and Arnott (2006) and using the wavelength from equation 1 gives a water depth of 35 m. In other words, at the time of Carstone deposition, in water depths >35 m the orbital velocity would have been <50 cm/s and hummocky or swaley cross-bedding would not have formed. This concurs with the findings of Dumas and Arnott (2006) who proposed that swaley cross-stratification typically forms in water depths between 13 to 50 m.

These calculations suggest that the Carstone cross-bedding formed on a storm-dominated shoreface, between fair-weather and storm wave base, in shallow water (30 to 40 m) where aggradation rates were low enough to allow preferential preservation of swales. This is supported by the absence of any fair-weather wave generated sedimentary structures, although accepting that these could have been destroyed by bioturbation.

Comparisons with Lower Greensand Group deposits

During the Aptian, marine connection between NE England and southern England was effected by the 'Bedfordshire Straits' (Figs 4 & 12), allowing seawater ingress into the Weald and Wessex Basins (Gale, 2012; Ruffell & Wach 1998b).

Sedimentary structures in these Lower Greensand Group sediments are mostly large-scale structures. The Aptian Woburn Sand Formation at Leighton Buzzard (Bedforsdhire Straits) contains cross bedding, between 1 and 10 m thick (Bristow, 1995), while the similar aged Hythe and Folkestone Formations further south feature large scale cross-bedding, with bioturbated clay drapes over the foresets (Allen, 1981). The general consensus is that these bedforms formed in strongly current swept seaways (Fig. 12) and in the 'Bedfordshire Straits' are indicating the migration of subaqueous sand waves and dune complexes in a southerly direction (Rawson, 1992). The Lower Albian Monk's Bay Sandstone Formation also contains large-scale (2-4 m thick) planar cross-bedding interpreted as tidally influenced sand-wave facies (Dike, 1972).

Cross bedding in the Carstone Formation differs from the English Lower Greensand deposits in three ways. (1) Mean cross-bed set thickness in the Carstone is ~15 cm, much smaller than values measured in the Folkestone Formation (~2 m; Allen, 1981), in the Woburn Sand Formation (1-10 m; Bristow, 1995) and in the

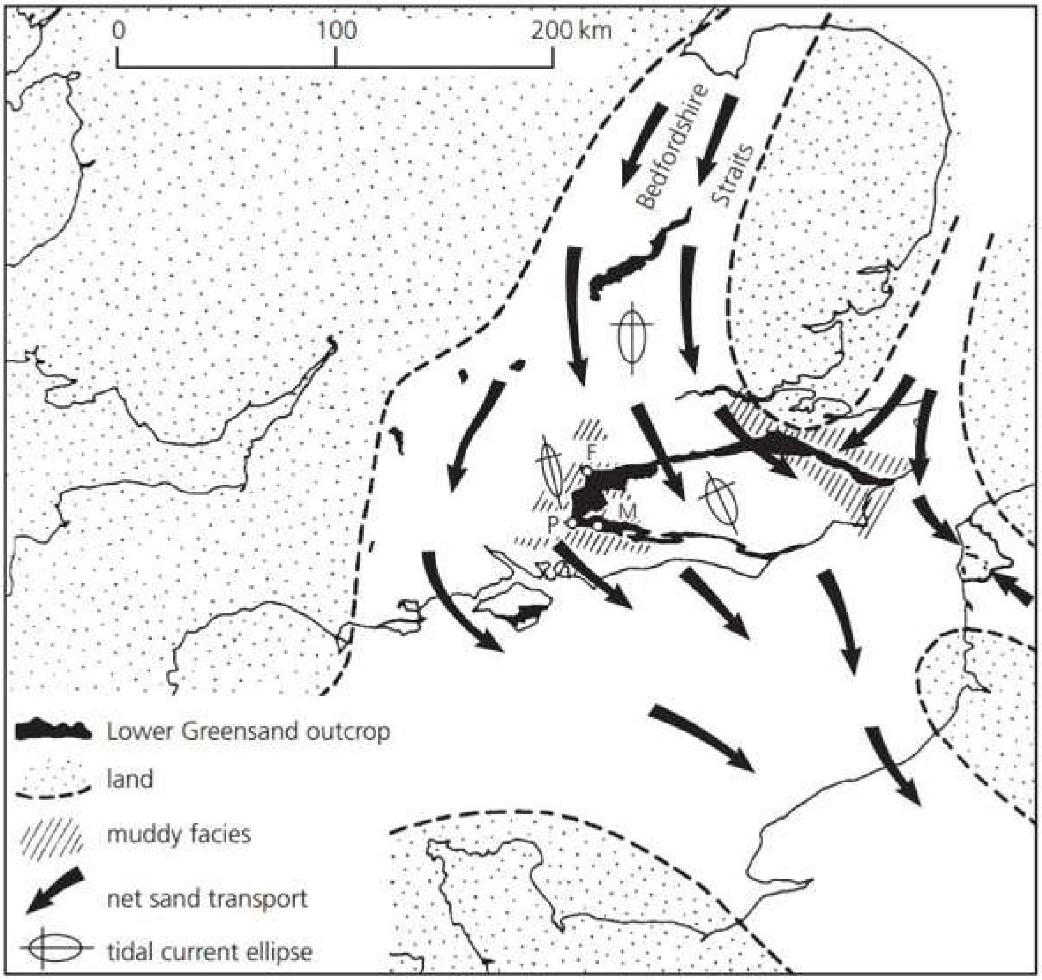


Fig. 12. Aptian-Albian paleogeography and sedimentation patterns in England (after Allen, 1982).

Monk's Bay Sandstone Formation (2-4 m thick; Dike, 1972). (2) The Carstone cross-bedding is seen only in medium to coarse-grained pebbly sand with complete absence of mud drapes on foresets. This contrasts with sand waves in the Lower Greensand Group, some of which are characterised by mud-draped bottomsets and foresets, linked to spring-neap depositional cycles (Allen, 1982). The mud layers, rarely exceeding a few mm thickness, are interbedded with sand on the lee side of dunes, believed to represent deposition during slack tides and periods of reduced flow. Allen (1982) suggests that the mud drape thickness and extent is a function of tidal asymmetry, eccentricity, current strength and sand wave asymmetry. The absence of mud drapes or indeed any muddy or silty beds in the Carstone

Formation suggest either a more proximal, shoreface setting, or heightened wave activity during which fine-grained sediment was widely dispersed. (3) The predominant foreset dip direction measured in the Carstone (this study; Gallois, 1994; Thomas, 1998) is toward the north or north-east, while most cross beds in the Lower Greensand deposits dip toward the south and south-east (Schwarzacher, 1953; Narayan, 1971).

Speculations on cross-bedding dip direction

The predominantly northward foreset dip directions in the Carstone are strikingly different to the predominantly southward foreset dips recorded in the Lower Greensand deposits (Fig. 12). Local 'estuarine geography' could have influenced Aptian-Albian current directions (cf. Hesselbo *et al*., 1990) and northerly Carstone palaeoflows at Blackborough End, 26 km S of Hunstanton have been interpreted as north/south trending tidal currents in an estuarine or incised valley fill environment (Thomas, 1988). However, the overall facies context at Hunstanton is one of open marine shelf conditions. We offer two interpretations to explain the directional data in an open shelf setting.

First, during a storm surge, wind forced currents cause coastal setup, creating a seaward pressure gradient (Fig. 13). During the storm surge ebbing water flows seawards but is deflected (in the northern hemisphere) to the right by the Coriolis force to form a geostrophic flow parallel to isobaths (Swift & Niedoroda, 1985; Fig. 13). On the modern Atlantic shelf in water depths between 10 and 20 m, Swift *et al*. (1979) recorded near seabed flow velocities up to 60 cm/s. These flows create ripples and dunes that would be preserved in the rock record as medium-scale cross-bedding. According to the paleogeography suggested by Anderton *et al.* (1979) and Allen (1982; Fig. 12), the eastern side of the 'Bedfordshire Strait' in Aptian-Albian times was bounded by a north-west facing shoreline. A storm surge relaxation current from the land deflected to the right could have resulted in an alongshore, north-eastern trending flow, potentially responsible for forming the Carstone cross-bedding. This interpretation allows for predominantly southward

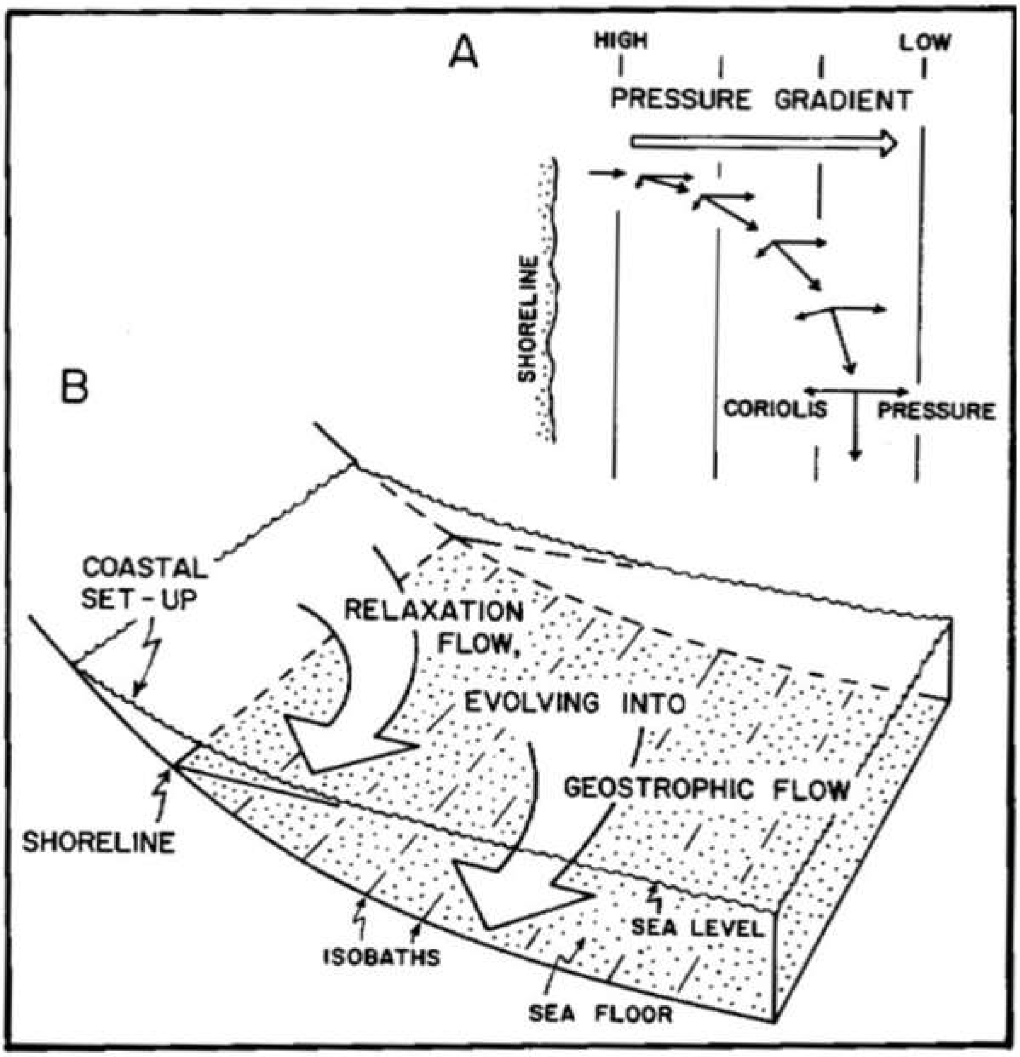


Fig. 13. Storm surge relaxation current deflected by the Coriolis force (after Walker, 1984).

ocean currents (Fig. 12), interrupted by episodic storms surges and the resulting storm relaxation currents. We also note that Dike (1972) interpreted parts of the time equivalent Monk's Bay Sandstone Formation (formerly Carstone) on the Isle of Wight as storm surge deposits.

Second, in experiments investigating combined flow conditions, Dumas *et al.* (2005) observed 'reverse', 30-40 cm high ripples that migrated upstream, forming upstream dipping laminae in the sediment. These bedforms formed under high oscillatory velocities (Uo 80-120 cm/s) and moderate unidirectional velocities (Uu < 15 cm/s) but were formed only in coarse-grained substrates where long oscillation periods (10.5 s) prevailed. These bedforms had an unusual migration direction because sediment was eroded on the short, steep, downstream side of the ripple and deposited on the longer, gently sloping, upstream side. The atypical migration direction was caused by cyclical motion in the oscillatory component of flow. When oscillatory motion is in the same direction as a unidirectional current, their effects combine to produce high flow velocity that causes intense scouring and a high concentration of suspended sediment at the base of the steep downstream side of the bedform (Dumas *et al.,* 2005). As the oscillatory flow stalls and reverses, this plume of sediment lifts and reverses, opposing the unidirectional current. The sediment cloud is then entrained upstream with bedform migration impaired by the downstream unidirectional current. Therefore, deposition occurs upstream of the flow separation point along the gently sloping upstream side of the bedform. In this way, reverse ripples migrate upstream, against a unidirectional flow (Dumas *et al.,* 2005).

In the Carstone depositional setting, oscillatory motion generated by waves combined with more unidirectional ocean currents from the north might plausibly have created upstream-dipping cross-bedding. The Carstone grain size range for the cross-bedded units is similar or slightly coarser than in the experiments of Dumas *et al.* (2005). The exact morphology and full height and of the Carstone cross-beds is not known due crest erosion, but >15cm (see above), so probably close to the 30-40 cm range of Dumas *et al.* (2005). The discussion above also allows for the few southward trending flow directions when storm currents relaxed, and may explain the southerly foreset dip directions mentioned by Versey & Carter (1926). We accept, however, that conditions for the formation of 'reverse' ripples are rarely met Dumas *et al.* (2005).

CONCLUSIONS

1. Swaley cross-stratification is identified in the Albian Carstone Formation at Hunstanton for the first time, corroborating an earlier identification by Thomas (1998) at nearby Sid George's Pit (Blackborough End).
2. The palaeoenvironmental significance of swaley cross-stratification is its generation below fair-weather wave base, but above storm wave base.

'Swales' are preferentially preserved under low aggradation rates (1 mm/min), conditions under which genetically related 'hummocks' are selectively eroded. Our calculations using published physical equations suggest the Carstone bedforms were generated on a storm dominated shoreface in 30 to 40 m water depth under combined flow with oscillatory and unidirectional components.

3. Palaeocurrent directions indicated by the Carstone cross-bedding foreset dip directions are northward, strikingly different to the predominantly southward palaeocurrents recorded in older Lower Greensand deposits of southern England. A storm surge relaxation current from the land deflected right by Coriolis forces could have resulted in an alongshore NE trending flow forming the Carstone cross-bedding. This interpretation allows for predominantly southward ocean currents interrupted by episodic storms surges and resulting relaxation currents.

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REFERENCES

ALLEN, J.R.L. 1981. Lower Cretaceous tides revealed by cross-bedding with mud drapes. *Nature*, **289**, 579-581.

ALLEN, J.R.L. 1982. Mud drapes in sand deposits: a physical model with applications to the Folkestone Beds (early Cretaceous, southeast England). *Philosophical Transactions of the Royal Society of London Series A*, **306**, 291–370.

ANDERTON, R., BRIDGES, P.H., LEEDER, M.R. & SELLWOOD, B.W. 1979. *A Dynamic Stratigraphy of the British Isles: a Study in Crustal Evolution*. Allen & Unwin, London. 301pp.

ANDREWS, J.E. 1983. A faunal correlation of the Hunstanton Red Rock with the contemporaneous Gault Clay, and its implications for the environment of deposition. Bulletin of the Geological Society of Norfolk, 33, 3-26

ANDREWS, J.E., KENDALL, A.C. & HALL, A. 2014. Microbial crust with *Frutexites*(?) and iron staining in chalks: Albian-Cenomanian boundary, Hunstanton, UK. *Geological Magazine*, **152**, 1-11.

BAAS, J.H. 1994. A flume study on the development and equilibrium morphology of current ripples in very fine sand. *Sedimentology*, **41**, 185-209.

BRISTOW, C. 1995. Facies analysis on the Lower Greensand using ground-penetrating radar. *Journal of the Geological Society of London*, **152**, 591-598.

CURRAY, J.R 1960. Sediments and history of Holocene transgression, continental shelf, northwest Gulf of Mexico. In: Shepard, F. P., Phleger, F. B. & van Andel TJ. H. (eds), *Recent Sediments, Northwest Gulf of Mexico*, 221-266, American Association of Petroleum Geologists, Tulsa, Oklahoma.

DeCELLES, P.G., LANGFORD, R.P. & SCHWARTZ, R.K., 1983. Two new methods of paleocurrent determination from trough cross-stratification. *Journal of Sedimentary Research*, **53**, 629-642.

DIKE, E.F. 1972. Sedimentology of the Lower Cretaceous Greensand of the Isle of Wight. Unpublished D. Phil thesis, University of Oxford.

DOTT, R.H. & BOURGEOIS, J. 1982. Hummocky stratification: Significance of its variable bedding sequences. *Geological Society of America Bulletin*, **93**, 663-680.

DUKE, W.L. 1985. Hummocky cross-stratification, tropical hurricanes, and intense winter storms. *Sedimentology*, **32**, 167-194.

DUKE, W.L., ARNOTT, R.W.C. & CHEEL, R.J. 1991. Shelf sandstones and hummocky cross-stratification: new insight on a stormy debate. *Geology,* **19**, 625-628.

DUMAS, S. & ARNOTT, R.W.C., 2006. Origin of hummocky and swaley cross-stratification - The controlling influence of unidirectional current strength and aggradation rate. *Geology,* **34**, 1073-1076.

DUMAS, S., ARNOTT, R.W.C. & SOUTHARD, J.B., 2005. Experiments on oscillatory-flow and combined-flow bed forms: Implications for interpreting parts of the shallow-marine sedimentary record. *Journal of Sedimentary Research,* **75**, 501-513.

EYERS, J. 1995. Correlation of the Lower Greensand Group (Woburn Sands and Carstone) of the Bedfordshire Province, England. *Cretaceous Research*, **16**, 385-413.

GALE, A.S. 2012. Early Cretaceous: rifting and sedimentation before the flood. In: Woodcock, N. & Strachan, R. (eds). *Geological History of Britain and Ireland* (2nd ed). 347-364, Wiley- Blackwell, Oxford.

GALLOIS, R.W.1994. Geology of the country around King's Lynn and the Wash. Memoir for 1:50000 geological sheet 145 and part of 129 (England & Wales). HMSO, London.

GALLOIS, R.W. 1988. The geology of the country around Ely. Memoir of the Geological Survey, Sheet 189. HMSO, London.

HARMS, J.C., SOUTHARD, J.B., SPEARING, D.R. & WALKER, R.G. 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Society of Economic Paleontologists and Mineralogists, Short Course Notes No. 2. 161pp.

HESSELBO, S.P., COE, A.L., BATTEN, D.J. & WACH, G.D. 1990. Stratigraphic relations of the Lower Greensand (Lower Cretaceous) of the Calne area, Wiltshire. *Proceedings of the Geologist's Association*, **101**, 265-278.

HOPSON, P. M. (2005). A stratigraphical framework for the Upper Cretaceous Chalk of England and Scotland with statements on the Chalk of Northern Ireland and the UK Offshore Sector. British Geological Survey Research Report RR/05/01. Keyworth, Nottingham.

HOPSON, P. M., WILKINSON, I P & WOOD, M A. (2008). The stratigraphical framework for the Lower Cretaceous of England. British Geological Survey Research Report RR/08/03. Keyworth, Nottingham.

HOPSON, P.M., WILKINSON, I.P., WOOD, M.A. & FARRANT, A.R. 2011. The Lower Albian (Lower Cretaceous) Monk's Bay Formation (formerly the Carstone) of the Isle of Wight: its distribution, litho- and biostratigraphy. *Proceedings of the Geologist's Association*, **122**, 816-830.

IMMENHAUSER, A. 2009. Estimating paleo-water depth from the physical rock record*.* *Earth Science-Reviews*, **96**, 107-139.

JOHNSON, H.D. & BALDWIN, C.T. 1996. Shallow Clastic Seas. In: Reading, H.G., (ed.) *Sedimentary Environments: Processes, Facies and Stratigraphy* (3rd ed.) Blackwell Science, Oxford, UK, 232-280.

LECKIE, D.A. & WALKER, R.G. 1982. Storm and tide-dominated shoreline in Cretaceous Moosebar-Lower Gates Interval – Outcrop equivalents of Deep basin gas trap in western Canada. *American Association of Petroleum Geologists' Bulletin,* **66**, 138-157.

LEEDER, M.R. 2011. *Sedimentology and sedimentary basins: from turbulence to tectonics* (2nd ed.) Wiley-Blackwell, Chichester. 768pp.

LI, M.Z. & AMOS, C.L., 1998. Sheet flow and large wave ripples under combined waves and currents: field observations, model predictions and effects on boundary layer dynamics. *Continental Shelf Research,* **19**, 637-663.

LINDHOLM, R.C., 1987. *A Practical Approach to Sedimentology*. Springer, Dordrecht. 160pp.

McKEE, E.D. & WIER, G.W. 1953. Terminology for stratification and cross-stratification. *Bulletin of the Geological Society of America*, **64**, 381-390

MITCHELL, S.F. 1995. Lithostratigraphy and biostratigraphy of the Hunstanton Formation (Red Chalk, Cretaceous) succession at Speeton, North Yorkshire, England. *Proceedings of the Yorkshire Geological Society,* **50**, 285-330.

MORSILLI, M. & POMAR, L. 2012. Internal waves vs. surface storm wave: a review on the origin of hummocky cross-stratification. *Terra Nova,* **24**, 273-282.

NARAYAN, J. 1971. Sedimentary Structures in the Lower Greensand of the Weald, England, and Bas Boulonnais, France. *Sedimentary Geology,* **6**, 73-109.

NICHOLS, G. 2009. *Sedimentology and Stratigraphy* (2nd ed). Wiley-Blackwell, Chichester, 419pp.

OWEN, H.G., 1995. The upper part of the Carstone and the Hunstanton Red Chalk (Albian) of the Hunstanton Cliff, Norfolk. *Proceeding of the Geologists' Association*, **106**, 171-181.

PRICE, G.D. & HARWOOD, E. 2012. Isotopic analysis of belemnites and brachiopods from the Cretaceous (Albian) Hunstanton Red Chalk Formation (Hunstanton, Norfolk, UK). *Proceedings of the Geologists' Association*, **123**, 479-485.

RAWSON, P.F. 1992. The Cretaceous. In: Duff, P. Mcld.D. & Smith, A.J. (eds) *The Geology of England and Wales*, 335-388, The Geological Society, Bath.

RUFFELL, A.H. & WACH, G.D., 1991. Sequence stratigraphic analysis of the Aptian-Albian Lower Greensand in southern England. *Marine and Petroleum Geology*, **8**, 341-353.

RUFFELL, A.H. & WACH, G.D. 1998a. Firmgrounds – key surfaces in the recognition of parasequences in the Aptian Lower Greensand Group, Isle of Wight (southern England). *Sedimentology,* **45**, 81-107.

RUFFELL, A.H. & WACH, G.D. 1998b. Estuarine/offshore depositional sequences of the Cretaceous Aptian-Albian boundary, England. In: Graciansky, P-D. de, Hardenbol, J., Jacquin, T. & Vail, P.R. (eds) *Mesozoic and Cenozoic Sequence Stratigraphy of European Basins*, 411-421. Society of Economic Paleontologists and Mineralogists Special Publication **60**, Tulsa Oklahoma.

SCHWARZACHER, W. 1953. Cross-bedding and grain size in the Lower Cretaceous sands of East Anglia. *Geological Magazine*, **90**, 322-330.

SHEPHARD-THORN, E. R. , MOORLOCK, B. S. P., Cox, B. M., ALLSOP, J. M. & WOOD, C. J. 1994. Geology of the country around Leighton Buzzard. *Memoir of the British Geological Survey,* Sheet 220. (England and Wales).

SWIFT, D.J.P., FREELAND, G.L. & YOUNG, R.A. 1979. Time and space distribution of mega-ripples and associated bedforms, Middle Atlantic Bight, North America Shelf. *Sedimentology,* **26**, 389-406.

SWIFT, D.J.P. & NIEDORODA, A. 1985. Fluid and sediment dynamics on Continental Shelves. In: Tillman, R.W. (ed.) *Shelf sands and sandstone reservoirs.* Society of Economic Paleontologists and Mineralogists Short Course Notes No. 13, 47-133.

THOMAS, C.J.S. 1998. Reservoir characterization of a shallow marine sandstone: The Lower Cretaceous Sandringham Sands (Leziate Beds) and Carstone formations, eastern England. *Petroleum Geoscience,* **4**, 215-219.

UNDERWOOD, C.J. & MITCHELL, S.F. 1999. Albian and Cenomanian selachian assemblages from North-East England. In: Unwin, D. (ed.) *Cetaceous Fossil Vertebrates* (Special papers in Palaeontology) No. 60, 9-56. Palaeontological Association.

VERSEY, H. & CARTER, C. 1926. The petrography of the Carstone and associated beds in Yorkshire and Lincolnshire. *Proceedings of the Yorkshire Geological Society*, **20**, 349-365.

WALKER, R.G. 1984. Shelf and shallow marine sands. In: Walker, R.G. (ed.) *Facies Models* (2nd ed), 141-170, Geological Association of Canada, Toronto, Canada.

WILKINSON, I.P. 2006. The Holostratigraphy of the Albian Stage (Lower Cetaceous) of the United Kingdom and its continental shelf. *British Geological Survey Research Report,* RR/06/01.

WOODS, M.A., WILKINSON, I.P., DUNN, J. & RIDING., J.B. 2001. The biostratigraphy of the Gault and Upper Greensand formations (Middle and Upper Albian) in the BGS Selborne boreholes, Hampshire. *Proceeding of the Geologists' Association*, **112**, 211-222.

YAGISHITA, K., ARAKAWA, S. & TAIRA, A. 1992. Grain fabric of hummocky and swaley cross-stratification. *Sedimentary Geology*, **78**, 181-189.

ZAR, J.H. 1981. Power of statistical testing: hypotheses about means. *American Laboratory*, **13**, 102-107.

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APPENDIX 1: ORIENTATION DATA

Structural data

Dip (degree)	Strike (degree)	Grid Reference (TF)
4 NE	116	67762 42311
4 NE	112	67807 42351
4 NE	102	67733 42294

Cross-bedding orientation

Sample	Dip (degree)	Dip direction (degree)	Strike (degree)	Grid Reference (TF)
1	19	64	334	67735 42291
2	25	58	328	67319 41513
3	35	66	336	67319 41513
4	16	24	294	67326 41544
5	26	131	41	67328 41551
6	20	209	119	67327 41556
7	21	229	139	67358 41714
8	21	52	322	67358 41714
9	23	59	329	67379 41770
10	30	93	3	67452 41897
11	18	343	253	67468 41922
12	25	351	261	67573 42048
13	25	57	327	67645 42136
14	16	71	341	67628 42111

Grain Orientation

Sample	Axis (degree)	Plunge (degree)
1	192	22
2	168	52
3	200	12
4	208	12
5	208	8
6	230	16
7	228	12
8	210	10
9	190	33
10	220	12
11	234	14
12	252	5
13	186	22
14	220	2
15	194	12
16	223	8
17	190	20
18	214	8
19	230	20
20	102	10

Sample	Axis (degree)	Plunge (degree)
21	140	16
22	209	21
23	194	24
24	212	13
25	206	8
26	194	10
27	16	12
28	20	6
29	47	14
30	334	10
31	56	5
32	47	10
33	9	21
34	19	19
35	3	22
36	28	24
37	54	18
38	8	7
39	30	16