maintained its jet-like characteristics further southeast that suggested by ALADIN at 1200 UTC. The change in wind direction observed is well reproduced by the model. Along leg BC, ALADIN evidenced the presence of a low level jet (corresponding to the 50-55 m/s) in good agreement with the measurements at 100 m (Figure A4). At 100 and 300 m, the winds were stronger in the middle of the leg, in accordance with ALADIN (i.e., the flight track is closest to the 20 m/s isobase near 4.6°W). The wind speed was also correctly reproduced by ALADIN.

Along leg AF, measurements suggested an increase of RH with longitude which is reproduced by ALADIN. At 1600 hPa, the RH increases from 25 to 45%, in fair agreement with the ARAT measurements, even though slightly underestimated by as much as 10%. Along leg BC, RH measurements indicate a slight increase with longitude from 40 to 55%. Simulated values are underestimated with respect to the measured ones.

[a] Temperature was observed to be relatively constant in the sheltered region (west of B). An important increase of the potential temperature (by 2 K) with the distance from the coastline was observed on legs BC (100 m ASL) and CR (300 m ASL), while on leg BF (300 m ASL), the potential temperature was observed to be constant. These fluctuations are well reproduced by the model.

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Djordjevic, D., P. Vukicevic, A. Jokic, L. T. Kostanos, Least-squares diabaticity method applied to derive turbulent fluxes over the ocean during the Surface of the Ocean, Fluxes and Interactions with the Atmospheric/Antarctic

Analysis of surface wind and roughness length evolution with fetch using a combination of airborne lidar and radar measurements

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[1] A combination of surface wind speed (SWS) and sea state variables, derived from quasi-simultaneous airborne lidar and radar measurements, made in the framework of the Flux, État de mer et Télédétection en Condition de fiéte variable (FETC) experiment, is used to analyze the evolution of surface roughness length, neutral drag coefficient, and friction velocity coefficient with fetch in the first hundred kilometers offshore over the Gulf of Lion, Western Mediterranean. The study focuses on the Tramontane/Mistral event documented in the afternoon of 24 March 1998. Particular attention is given to SWS derived from lidar measurements. The SWS retrieval methodology developed and validated at the Near Infrared ocean conditions by Flamant et al. [1998] has been modified to account for the specificity of the coastal Mediterranean environment (complex mixture of continental and maritime areas). It is further shown that, based on a formulation of the dimensionless roughness dependence with wave age, airborne lidar and radar measurements can be combined to provide insight into the evolution with fetch of roughness length, neutral drag coefficient, and friction velocity. Four distinct sea state regimes over a distance of 100 km could be identified from the remotely sensed variables obtained with this novel approach in connection with atmospheric forcing. The dependence of lidar-derived drag coefficient with lidar-derived SWS for the four regimes was found to be remarkably consistent with the relationship derived from the buoy measurements. Finally, lidar-derived friction velocities were found in good agreement with the buoy and in situ aircraft measurements.

INDEX TERMS: 3359 Meteorology and Atmospheric Dynamics: Oceanatmosphere interactions (0312, 4504); 3360 Meteorology and Atmospheric Dynamics: Remote sensing; 4506 Oceanography: Meteorology, Oceanography, and Atmospheric Dynamics: Methods: Remote sensing.

related to the “roughness” of the waves and depends on atmospheric wind speed, atmospheric stability, locally generated wind-waves (i.e., wave size, shape and phase velocity) as well as swell (e.g., Huang and Sheddin, 1988; Pickers and Mohri, 1997 and references therein; Drennan et al., 2003).

[1] To this day, most of the process-oriented investigations dedicated to the analysis of momentum flux in limited fetch conditions have been conducted from still and slowly moving sea-borne platforms, i.e., buoys, ships, offshore towers or a combination of platforms (e.g., RASEX (Riso Air Sea Experiment) [Mahr et al., 1996] or WAVES (Water-Air Vertical Exchange Study), SWADE (Surface Wave dynamics Experiment) and HEXOS (Humidity Exchange over the Sea)), see a brief overview of these experiments in the work of Drennan et al. [2003]. In such studies, the variety of the fetch conditions encountered at the platform results essentially from the variety of wind conditions observed in the course of an experiment. Direct measurement of momentum fluxes in the MABL (by eddy-correlation or inertial dissipation techniques) require statistically representative sets of data (i.e., long integration times) and assume quasi-stationary conditions at the point of measurements.

[2] In contrast, the study of momentum transfer evolution with fetch in offshore conditions has been a largely neglected area of research, until now, due to the fact that momentum flux are difficult to measure from airborne platforms. Concerning these fluxes, representativity and uncertainty issues arise from: (1) the difficulty of making measurements in the shallow surface layer (i.e., the “constant (flux)” layer) with an aircraft and (2) the technique employed to measure the fluxes (i.e., an eddy correlation technique). The eddy correlation technique requires thermodynamical variables to be accumulated over a statistically representative sample (i.e., 20–25 km) in which case the assumption of quasi-stationary conditions is not valid. Moreover, the horizontal resolution of such measurements (20–25 km) far exceeds the resolution need to examine surface related processes.

[3] The study of momentum flux evolution with fetch, in the “Lagrangian” sense, is highly desirable for a better comprehension of momentum transfer at the mesoscale and improving coupled ocean-atmosphere circulation models in coastal regions. Remote sensing of surface variables from airborne platforms appears to be the natural link between local measurements from buoys and ships and those retrieved at the mesoscale from spaceborne platforms and simulated by atmospheric models. The parameterization of surface flux from remotely sensed variables needs to be developed.

[4] In this paper, we propose a novel approach combining airborne radar and lidar measurements to determine, in a coastal region, the evolution with fetch of variables which are key to the understanding and parameterization of momentum fluxes, i.e., roughness length and drag coefficient. The capability of airborne radar ocean wave spectrometers to provide precise, high spatial resolution measurements of surface wind speed, significant wave height and wave peak frequency has been demonstrated repeatedly [Jackson et al., 1985; Hussey et al., 1992, 1995; Banner et al., 1999; Pettersson et al., 2003]. More recently, airborne lidars have shown promising potential for providing accurate highly resolved surface wind speed (SWS) over the ocean [e.g., Bjothon et al., 1983; Flamant et al., 1998]. As discussed by these authors, sea surface reflectance can be inferred from airborne radar lidar measurements by analyzing the interaction of the laser pulse with the sea surface. Provided that atmospheric effects (variability in both aerosol composition, aerosol concentration and relative humidity) and oceanic effects (the contribution of the submarine reflectance) can be corrected for, the sea surface reflectance can be related to the surface mean square slope produced by capillary and gravity waves riding on longer waves. Surface mean square slope can then be interpreted in terms of SWS using the comprehensive work of Cox and Munk [1954]. This technique has also been applied successfully to spaceborne lidar data acquire during the Lidar In-space Technology Experiment [Menzie et al., 1998].

[5] In this paper, we use a combination of airborne lidar and radar measurements acquired in the framework of the Flux, Etat de mer et Télédétection en Condition de FETCH variable (FETCH) experiment [Hussey et al., 2003, 2002], to analyze the evolution of sea state variables with fetch in the first hundred kilometers offshore over the Gulf of Lion (Gol), Western Mediterranean. The study focuses on the Tramontane/Mistral event documented in the afternoon of 24 March 1998. The Tramontane and the Mistral are low-level, orography-induced, cold-air out­breaks over the Gol, blowing offshore of Narbonne and Arles, France, respectively (Figure 1). They are frequently observed to extend several hundreds of kilometers from the coast, bringing cold and dry continental air over the warm Western Mediterranean, and hence, generating intense heat air-sea exchanges. They are one of the primary cause of storms over the Mediterranean, between Italy and the Balearic Islands [Trigo et al., 1999; Campus et al., 2000].

[6] The analysis also relies on the measurements at sea performed by a moored buoy and a ship specially deployed for the experiment, as well as measurements from the TOPEX altimeter. This paper addresses several questions, including the following:

1. Can SWS be inferred from lidar measurements over the sea in a coastal environment (complex mixture of continental and maritime aerosol; turbulent, productive waters)?
2. What is the influence of developing waves on the lidar-derived surface mean square slope and related SWS?
3. How does spatial evolution of sea state variables (SWS, roughness length and drag coefficient) with fetch derived from airborne measurements compare with the temporal evolution of these variables derived from a moored buoy?

[7] In section 2, we present the synoptic situation and the FETCH operations on 24 March 1998. In section 3, we describe the methodology used to derive sea state variables from lidar (SWS) and a combination of lidar and radar data (roughness length and drag coefficient) over coastal waters. In section 4, we discuss the evolution of surface mean square slope measured by airborne lidar, airborne radar (C-Band) and spaceborne radar (C-Band and Ku-Band). We compare the evolution of lidar-derived SWS and roughness
length with measurements made from other platforms (i.e., sea-borne, ship-borne, airborne and spaceborne platforms). We also discuss the spatial evolution of the momentum flux drag coefficient derived from airborne measurements. Concluding remarks are in section 5.

2. The 24 March 1998 Case of Fetch: Synoptic Situation and Operations

[i] In addition to the means specially deployed for FETCH, Météo-France provided daily forecasts of the meteorological situation made with the operational model ALADIN. The forecasts on 24 March 1998 have been validated against buoy, ship and aircraft in situ data [Flamant, 2003]. These comparisons confirmed that the thermodynamical conditions prevailing over the Gol during the afternoons airborne operations (see below) were well represented by the 1800 UTC ALADIN forecast. In turn, the ALADIN forecasts were used by Flamant (2003) to analyze the rapid evolution of the synoptic situation on 24 March 1998. As these forecasts are also utilized in this study, we first briefly present the ALADIN model.

[ii] ALADIN is a spectral limited area model, taking its boundary conditions from the global model ARPEGE of Météo-France, which covers a domain of 2739 km x 2739 km (centered on France). The horizontal resolution is approximately 10 km, with 31 levels on the vertical, the highest level being at 5 hPa and the lowest level at approximately 17 m above ground/sea level. The surface layer and planetary boundary layer fluxes are computed on the lowest level using a modified version of the scheme developed by Louis et al. [1981]. The 3D-var data assimilation provides two analyses per day (0000 and 1200 UTC) but no associated first guess. Forecasts are available every 3-h, i.e., at 0000, 0300, 0600, 0900, 1200, 1500, 1800, 2100 and 2400 UTC. 3-h integrated surface turbulent fluxes are computed from 12-h forecasts. Additional information can be found in the work of Eymard et al. [2003]. The forecasts products are: temperature, humidity and wind distributed on 16 pressure levels between 1000 and 100 hPa. Sea level pressure and surface turbulent fluxes are provided at the lowest level, i.e., 17 m ASL. Near surface extrapolated fields (at 10 m ASL for the wind and 2 m ASL for temperature and humidity) are derived assuming a neutral
profile. The SST in the model is issued from Reynolds’ climatology and yields characteristic spatial and temporal scales of about 2° and 8 days, respectively.

### 2.1. Synoptic Situation

[i] Based on ALADIN forecasts, Flamant [2003] has shown that the monsoon nature of the Tramontane/Mistral wind regime over the Gol, on 24 March 1998 was controlled by the multistage evolution of an Alpine lee cyclo LCS over the Tyrrhenian Sea (between Sardinia and continental Italy). In the early stage, the Tramontane flow prevailed over the Gol. As the low deepened, the prevailing wind regime shifted to a well established Mistral which peaked around 1200 UTC. In the afternoon, the Mistral was progressively disrupted by a strengthening outflow coming from the Ligurian Sea in response to the deepening low over the Tyrrhenian Sea and the channeling induced by the presence of the Apennine range (Italy) and the Alps. In the evening, the Mistral was again well established over the Gol as the depression continued to deepen but moved to the south-east, reducing the influence of outflow from the Ligurian Sea on the flow over the Gol.

### 2.2. Operations

[ii] On 24 March 1998, measurements of the mean and turbulent properties within the ABL over the Gol were made from four platforms: two aircraft (the ARAT—Avion de Recherche Atmosphérique et Télédétection—and the Merlin IV), the Research Vessel Atalante and the Air-Sea Interaction Ship (ASIS) buoy. The operations were coordinated with TOPEX descending overpass 146 (1844 UTC).

[iii] The Research Vessel Atalante had balloon launching capability and carried an instrumented mast for mean and turbulent measurements at a height of 17 m above the sea surface. The ASIS buoy [Braun et al., 2000] made measurements of mean and turbulent atmospheric variables 7 m above the air-sea interface as well as wave directional spectra. The ARAT and Merlin IV were equipped with standard in situ sensors as well as sensors dedicated to the surface-air interaction (cloud condensation nuclei counters). The ARAT and the Merlin IV also carried the differential absorption lidar LEANDRE 2 [Bramenou et al., 2000a, 2000b] and the real-aperture airborne radar RESSAC (Radard pour l’Etude du Spectre de Surface par Analyse Circulaire) [Hauter et al., 1999], respectively. RESSAC is a C-Band radar, with a large antenna 

### 3. Methodology for Sea State Variable Retrieval

[i] At this point, it should be noted that the airborne lidar LEANDRE 2 was not developed for the primary purpose of measuring SWS [LEANDRE 2 was developed for the purpose of analyzing surface-atmosphere exchanges through the ABL]. This implies careful lidar calibration and inversion as described in Appendix A and B. A lidar system specifically developed for the purpose of measuring SWS (e.g., a system with longer wavelength and multiple look angles) would alleviate some of the limitations of the calibration and inversion approach described in this paper.

[ii] To analyze the lidar data acquired during FETCH, the SWS retrieval methodology developed and validated for open ocean conditions by Flamant et al. [1998] has been modified to account for the specificity of the coastal Mediterranean environment, namely: (1) coastal aerosol composition (to correct for atmospheric effects), (2) foam at the sea surface and air bubbles within the water and (3) submarine reflectance enhancement at 732 nm caused by the presence of Chlorophyll, inorganic suspended particles from terrestrial origin, yellow substances (in the Rhone river plume) or even mineral dust (deposited at the surface during Sahara dust outbreaks). Note that the foam and subsurface water contributions to the lidar signal can be separated in two distinct components: (1) one related to the sunlight diffusion by the subsurface water and/or foam in the direction of the lidar telescope and (2) one related to the interactions between the laser pulse and the subsurface water and/or foam within the laser footprint. It is argued that solar radiation diffused by the subsurface water (including that of suspended particulate) and/or foam in the direction of the lidar telescope is removed through “background” light correction applied to the lidar data (see Appendix A). Solar-stimulated Chlorophyll fluorescence is also argued not to be a factor at 732 nm.

[iii] Atmospheric effects (variability in both aerosol composition, aerosol concentration and relative humidity) and their influence on sea surface reflectance (SSR) retrievals...
flament at al.: surface wind speed and roughness length

3.1. Lider-Derived Surface Meaning Square Slope

where \( S^2 \) is the total wave MSS variance. In this expression, the effects of laser pulse shape and divergence have been neglected as contributing very weakly to our measurement configuration [Gardner et al., 1983]. The SWS through a linear relationship [Cox and Monk, 1954].

The methology for retrieving surface MSS. This is discussed below (section 3.1). The TOPEX and RESSAC are presented in Appendix C. Besides the contribution of waves, it is argued that laser radiation backscattered by whitecaps and the light backscattered from the first of the terms in equation (1). Finally, the contribution from foam depends on the area covered by whitecaps. In the present experiment, the output energy was adjusted so that lidar surface returns were not saturated for small fractional whitecap coverage (low albedo). For a large fractional whitecap-coverage, the high albedo associated with foam could induce a saturated surface return and a non-linear response of the lidar system detector. The only efficient way to screen out this contribution is to discard the lidar reflectivity profiles with saturated surface echoes. This is the case for the lidar data used in this paper. When the fractional area covered by whitecaps is small, the foam related reflectance needs to be assessed. It is argued that laser radiation backscattered by foam can be accounted for based on the work of Koepke et al. [1998] and a power law relationship between wind speed and coverages as discussed below. The contribution of waves was derived from the Rossby–Sutcliffe equations [Cox and Monk, 1954], the surface MSS statistics have been observed to be constrained by the light surface. In clean ocean water (water molecules, organic and inorganic matter), this is generally less than 1% at 752 nm. In coastal regions, this term cannot necessarily be neglected in the near-infrared part of the spectrum due to the presence of suspended particulate (yellow matter, mineral dust or even anthropogenic aerosols such as soot). Furthermore, the pro-

3.2. Lider-Derived Surface Wind Speed

\[ \rho = W_{SWS} + (1 - W_{SWS}) (1 - W_{Foam}) + (1 - W_{Foam}) \rho_{Foam} \]  

\[ \rho = \frac{\rho_{Foam}}{1 - W_{Foam}} \]  

where \( \rho_{Foam} \) is the total wave MSS variance. In this expression, the effects of laser pulse shape and divergence have been neglected as contributing very weakly to our measurement configuration [Gardner et al., 1983]. The SWS through a linear relationship [Cox and Monk, 1954].

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stability of the atmospheric surface layer departed from neutral stability (i.e., the observations of Cox and Monk [1954] were made in slightly positive stability conditions). Show and Churnside [2001] reported observations in the negative stability regime (air temperatures cooler than water temperatures), demonstrating increasing relative MSS (their MSS normalized by the Cox and Munk values) with increasing negative stability. The data discussed here was acquired in cold-air outbreak conditions for which the negative stability is expected to be important. Accounting for atmospheric stability, the wave MSS is given by [Show and Churnside, 2001]

\[
R_i = (g\Delta T_w + 0.38U^2)^{1/7},
\]

where \( R_i \) is the reduced Richardson number given by

\[
R_i = (g\Delta T_w + 0.38U^2)^{1/7},
\]

with \( g \) the gravitational constant, \( \Delta T_w \) the air-sea temperature difference (°C), \( T_w \) the sea temperature (°C), and \( U \) the wind speed measured at height \( h (10 \text{ m}) \). Note that equation (3) and equation (4) are the same for \( R_i = 0.15 \) which corresponds approximately to the stability regime under which were acquired the wave MSS reported by Cox and Monk [1954]. The Richardson number along leg AP was estimated using ALADIN forecasts at 1800 UTC (see values of the above cited parameters in Table 1). It was estimated to range between \(-0.13 \) and \(-0.065 \), yielding a correction term for the wave MSS between \( 1.6 \) and \( 1.8 \) (i.e., for a given SWS, the wave MSS is a factor \( 1.7 \pm 0.1 \) larger than that reported by Cox and Monk [1954]).

The final step, which consists of deriving \( U \) from lidar-derived \( \bar{p} \), is not trivial because \( \bar{p} \) is the sum of two compensating terms, \( p_h \) and \( p_p \), that both depend on \( U \) as

\[
p_h = (1 - W) \frac{p_u}{f(\bar{p})},
\]

and the modeled dependence of foam reflectance calculated as [Koopke, 1984]

\[
p_p = dU/dW,
\]

where the fractional area covered by whitecaps \( W \) can be expressed as a power law of \( U \). Hence, the reflectance due to the slope distribution of the capillary-gravity waves (\( \phi_u \)) decreases with increasing SWS, whereas the reflectance due to the slope distribution of the capillary-gravity waves (\( \phi_p \)) increases with increasing SWS.

[39] Figure 2 shows the evolution of the SSR (i.e., \( p_h + p_p \)) as a function of \( U \). It also illustrates the sensitivity of the SSR to (1) atmospheric stability conditions and (2) the relationship between SWS and whitecap coverage. Here, we have considered two types of stability conditions: the slightly stable conditions experienced by Cox and Monk [1954] (i.e., \( \Delta T_w \) given by equation (3)) and the unstable conditions experienced on 24 March 1998 (i.e., \( \Delta T_w \) given by equation (4)). Two types of whitecap coverage dependence on SWS have also been considered: the relationship established on 24 March 1998 using a combination of digital images of the sea surface made from the Research Vessel Atlantica and wind speed measurements made on the ship, based on the technique described by Dupuis et al. [1993], yielding

\[
W = 1.57 \times 10^{-6} U^{0.16}
\]

and the relationship of Monahan [1986]

\[
W = 3.84 \times 10^{-6} U^{0.41}
\]

[41] In Figure 2, the 4 curves correspond to:

1. Curve 1: the wind-waves contribution representative of the average unstable conditions experienced on 24 March 1998 along leg AF, i.e., for a negative stability correction term equal to \(-1.7 \), and the foam related reflectance computed using the whitecap coverage dependence on surface wind speed established on 24 March 1998.

2. Curve 2: same as curve 1, but for the slightly stable conditions encountered by Cox and Monk [1954].

3. Curve 3: same as curve 1, but for the whitecap coverage dependence on surface wind speed established by Monahan [1986],

4. Curve 4: same as curve 2, but for the whitecap coverage dependence on surface wind speed established by Monahan [1986].

[42] For the range of SWSs considered here (i.e., \( 0 - 30 \text{ m s}^{-1} \)), the foam related reflectance obtained from equation (8) is negligible (curves 1 and 2), and the total reflectance

<table>
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<td>4.6</td>
<td>11.1</td>
<td>13.3</td>
<td>12.1</td>
<td>-0.076</td>
<td>0.45</td>
</tr>
<tr>
<td>42.5</td>
<td>4.6</td>
<td>10.0</td>
<td>13.3</td>
<td>12.2</td>
<td>-0.080</td>
<td>0.45</td>
</tr>
<tr>
<td>42.4</td>
<td>4.7</td>
<td>8.9</td>
<td>12.5</td>
<td>12.2</td>
<td>-0.114</td>
<td>0.45</td>
</tr>
</tbody>
</table>

\[ *R_i \text{ is the Air Temperature (°C), } T_w \text{ is the Sea Temperature (°C), } R_i \text{ is the Reduced Richardson Number, } U \text{ is the Wind Speed Measured at } 10 \text{ m ASL, } \rho \text{ is the Freesurface Velocity and } L \text{ is the Monin-Obukhov Length.} \]
The reflectance due to whitecaps can be obtained unambiguously from total reflectance curve shown in Figure 2 because we only have to consider the leftmost branch of the curve. Nevertheless, SWS values cannot necessarily be derived for each lidar reflectance measurements because there exist a minimum SWS value (defined by \( \frac{\Delta U}{U} = 0 \)) associated with the SWS threshold value. In the case of curve 3, no SWSs can be inferred from lidar-derived surface reflectance values less than 0.044 (i.e., 61 out of 288 reflectance retrievals along leg AF). In the case of curve 4, no SWSs can be inferred from lidar-derived surface reflectance values less than 0.066 (i.e., 249 out of 288 reflectance measurements). In this case, considering the uncertainty on lidar-derived SWS (of the order of 10%, see Appendix B), even fewer SWSs can be inferred from lidar data. [3] In Figure 3, we show the evolution with fetch of the lidar-derived SWSs obtained from the 4 curves discussed above. Also shown is the SWS measured at 7 m ASL by the ASIS buoy at 1634 UTC. One may note that lidar-derived SWSs are larger in the slightly stable case than in the
unstable case. This is due to the higher modeled surface reflectance obtained in stable conditions as illustrated by curve 2 in Figure 2. The agreement between the ASIS measurements and the SWS retrievals in unstable conditions (curve 1) is very good. Hence, there is evidence that the negative stability conditions should be accounted for. Ignoring the stability correction term (i.e., assuming the conditions to be quasi-neutral), the SWS would be 0.7 m s\(^{-1}\) larger on average, along leg AF. In the remainder of this paper, the SWS have been derived from lidar measurements accounting for the unstable atmospheric conditions.

In unstable conditions, the choice of the fractional whitecap coverage dependence on \(U\) has an impact on the lidar retrievals for \(U > 12\) m s\(^{-1}\) only (see the difference between curve 1 and curve 3). For lidar-derived surface reflectance values greater than 0.044, the larger SWSs obtained from curve 3 than from curve 1 for \(U\) between 10 and 16.8 m s\(^{-1}\) result from the weaker dependence of \(\rho\) on \(U\). Given the uncertainty on lidar-derived SSR (Appendix B) and given that the ASIS derived SWS at the time of the ARAR overflight (1634 UTC) falls between the 2 curves, we have no mean of assessing which of the 2 \(\rho\) versus \(U\) relationships presented earlier best represents the conditions of the experiment. In the following, we have chosen to use the relationship derived on 24 March 1998 (equation (8)) because it enabled the retrieval of a SWS value for each of the lidar-derived SSR and because the lidar-retrievals were in slightly better agreement with the ASIS measurement. The uncertainty introduced (i.e., by using equation (8) rather than equation (9)) increases exponentially with SWS, but remains below 20% for SWS less than 15 m s\(^{-1}\). It is less than 3% for SWS less than 10 m s\(^{-1}\).

The uncertainty on the so-obtained SWS is taken as the quadratic sum of three terms: (2) the error associated with the uncertainty on the stability correction term (1.7 ± 0.1) which is on the order of 12%, (2) the error on the near surface backscatter coefficient (10%), and (3) the uncertainty resulting from the natural variability associated with the data, which is calculated as the standard deviation of the SWS computed for each individual profile (prior to the lidar slits average). The natural variability was observed to be quite significant, on the order of 20%. Hence, the total uncertainty associated with the SWS retrievals is equal to 25%.

3.3. Momentum Flux Related Variables From Lidar and Radar Measurements

As mentioned in the introduction, very little has been done on the study of momentum transfer evolution with fetch in offshore conditions. In this paper, we would like to take
AND ROUGHNESS for different atmospheric stability labels the retrieval of a...roughness coefficients determined from those data alone. [40] The neutral drag coefficient is then obtained from lidar/radar measurements as

$$C_{D0} = \frac{s}{2u^*}$$

and the friction velocity is obtained as

$$u^* = \sqrt{C_{D0}f_{SW}}$$

4. Results

In this section, we discuss the evolution of MSS, SWS and SRL with fetch as obtained from the different platforms along leg AF. The comparison was complicated by (1) the fact that the measurements from the different platforms were not made simultaneously in the same location and (2) the nonstationary nature of the Mistral flow over the Gulf of Genova (Figure 4a). As a result, the Mistral outbreaks over the Gulf on 24 March 1998 (see section 2). Hereafter, we begin with a description of the spatiotemporal evolution of the atmospheric forcing and its impact on sea state.

4.1. Nonstationary Aspect of the Mistral Flow Between 1500 and 2100 UTC

[45] Nondirectional spectra from 24 March 1998 are presented in the ALADIN wind field at 10 m ASL over the Gulf of Genova at 1500, 1800 and 2100 UTC. An important feature associated with cold-air outbreaks over the Gulf is observed in the form of barriers of weaker winds (sheltered region) shown as regions where the wind is less than 10 m/s.

[44] At 1200 UTC, a region of winds less than 10 m/s was observed in the eastern part of the domain, which corresponds to the sheltered region (associated with the Maritime Alps) separating the Mistral from the outflow from the Gulf of Genova (Figure 4b). As a result, the Mistral appeared to lose its characteristics over the sea, where the largest winds associated with the mistral were observed over the continent, in the Rhone Valley. At 1800 UTC, the Tramontane flow was observed to be shifted to the southwest as the Mistral began to intensify again. The influence of the sheltered region to the east of the Mistral also appeared to weaken. The sheltered region separating the Mistral and the Tramontane was very narrow at that time (Figure 4b). The cold and jet-like characteristics of the Mistral were only established after 2100 UTC, the Tramontane flow was well established again. It had a more marked northerly direction than at 1200 UTC due to the southward displacement of the Alpine lee cyclone (Figure 4c) [Flamant, 2001].
calculated as the distance from the coast assuming the wind to be northerly as shown by ALADIN model (see the location of the spectra in Figure 4). The spectra behaved as expected for limited fetch situations, i.e., they exhibited an increase of the energy and decrease of the peak frequency for increasing fetch distance. The peak frequency evolved with fetch as predicted by the JONSWAP relationship [Hasselmann et al., 1973] for a wind speed of the order of 15 m s⁻¹, although the significant wave height increase more rapidly than with the JONSWAP relationship.

Analysis of the wave directional spectra showed a marked swell component in the southern part of the track. For the same four locations, the normalized energy density (with respect to the peak) are plotted with contour levels in Figure 6. The ambiguity in the direction of RESSAC spectra was resolved with the output of the numerical sea state prediction model, WAMMED coupled with ALADIN 10 m ASL winds. WAMMED is the “Mediterranean version” of the operational WAM wave prediction model at ECMWF (European Center for Medium-range Weather Forecasts) [Bidlot, 2001, available at http://www.ecmwf.int/publications/newsletters/list.html] which provides routinely the directional wave spectrum at a 0.25° x 0.25° resolution, and uses as input the wind field of the ECMWF global atmospheric model IFS (Integrated Forecast System). Here a research version of WAMMED was used: the resolution was increased to 0.083° x 0.083° in latitude/longitude and the wind-forcing was provided by ALADIN at the same resolution. The directions refer to waves going to. For the locations 1 and 2, directional spectra were bimodal, the two peaks of similar frequency propagating on both sides of the wind direction (northerly). These waves are recognized as being generated by the action of the Mistral. At locations 3 and 4, three wave components were observed. One was propagating to the south and was identified as a Mistral generated wave train. The second component was propagating at lower frequency to the south-east and resulted from the action of the Tramontane blowing to the west of the region of operation. The last component with the lowest frequency was propagating to west-southwest. These waves were related to swell advecting from the Ligurian sea (as discussed above on the basis of ALADIN forecasts, Figure 4) and contributed up to one third of the total wave energy at the location 4.

In the northern part of the track (north of 42.75°N), sea-state is dominated by the Mistral flow and is limited in...
advecting to west-southwest. Theses basis of ALADIN forecasts, SOuth-east COl11ponent field of the ECMWF global blowing to the west of the French operational AUf!ffiC)on one third of the total wave Was identified as a track (north of 42.75°N), was propagat­

Waves growing not only due to the wind wave evolution with the fetch laws, but also with the swell contribution. Only 80% of the significant wave height is related to wind sea in that region.

4.3. Evolution of Radar, Lidar, and TOPEX Mean Square Slopes With Fetch

(continued) The evolution of lidar- and radar-derived (C-Band and Ku-Band) MSS as a function of the distance to the coast on leg AF is shown in Figure 7. Radar (TOPEX and RESSAC) MSS are obtained according to the methodology presented in Appendix C. The lidar-derived MSSs are obtained from equation (2) assuming the foam contribution can be neglected in the 24 March 1998 case (see discussion in section 3.2). The largest (smallest) MSSs were obtained from lidar (radar C-Band) measurements. This is related to the wave number interval over which the energy spectrum is integrated to yield the MSS (i.e., \( S(k) = \int_{k_{min}}^{k_{max}} \omega^2(k, S) dk \), where \( k \) is a function describing the evolution of the wave energy as a function of wave number \( k \). The wavelength interval to which instruments are sensitive is defined by the wavelength of operation (lower limit) and the footprint (upper limit). The wavelength intervals (in m) of LEANORD 2, TOPEX Ku-Band, TOPEX C-Band and RESSAC are [0.01, 30], [0.08, 1000], [0.2, 1000] and [0.2, 500], respectively. The corresponding wave number intervals (in m\(^{-1}\)) are [0.7, oo], [6 \times 10^{-4}, 80], [6 \times 10^{-3}, 30] and [0.01, 30], respectively. Hence, the larger MSS derived from lidar measurements reflect the fact that the energy spectrum is integrated (1) over a wider wave number range and (2) in a region of the spectrum sensitive to \( S\). Due to the smaller footprint, lidar-derived MSS exhibits a greater variability than its radar counterparts. Nevertheless, the 4 curves exhibit very similar trends as a function of latitude: first, a MSS steadily increasing with decreasing latitude (increasing distance from the coast) between the coast and 43°N, then a sharp decrease in MSS followed by a relatively constant MSS (43–42.6°N) and finally a MSS increasing again with the distance from the coast (42.6–42.4°N). The increase in the 42.6–42.4°N region is thought to be related to the presence of swell coming from the Gulf of Genova and cannot solely be interpreted in terms of SWS (i.e., the influence of swell has been discarded from the data used by Cox and Munk [1954] to establish the empirical relationship given by equation (3)). Radar data also showed that the MSS increased before stabilizing further away from the coast.

4.4. Evolution of Surface Wind Speed With Fetch

The evolution of lidar-derived SWS as a function of the distance from the coast on leg AF (12 m ASL, between 1620 and 1640 UTC) is illustrated in Figure 8. It is compared to the SWS simulated by ALADIN (10 m ASL, at 1800 and 2100 UTC), and to measurements made by ASIS and on­

the board of the Atalante (10 m ASL, at 1600 and 1700 UTC) as well as by TOPEX (10 m, at 1844 UTC) and in situ measurements performed from the ARAT (between 1750 and 1850 UTC) at 100 m ASL and at 300 m ASL (leg FB and leg BA, respectively, see Figure 1).
Figure 4. (continued)

RESSAC date: March 24 1998

RESSAC data obtained from RESSAC at four locations:

- **Location 4:** 42.28N, 4.84E
  - SWH (m) = 2.92
  - Peak frequency (Hz) = 0.131

- **Location 3:** 42.53N, 4.64E
  - SWH (m) = 2.63
  - Peak frequency (Hz) = 0.146

- **Location 2:** 42.78N, 4.44E
  - SWH (m) = 2.16
  - Peak frequency (Hz) = 0.162

- **Location 1:** 43.02N, 4.24E
  - SWH (m) = 1.74
  - Peak frequency (Hz) = 0.175

Figure 5. One dimensional spectra obtained from RESSAC at four locations: (1) near the shore in thick solid line, (2) in dotted line, (3) in dash-dotted line, and (4) in open sea in thin solid line. The energy density spectrum is plotted in log-log scale. The horizontal axis is for the frequency (Hz).
Topex SWS are estimated from the normalized radar cross section according to a model function defined by the modified Chelton and Wentz algorithm [Witter and Chelton, 1991]. The model function used here is a least squares fit of a fifth order polynomial to the tabular model of Chelton and Witter (1991). The comparison was made regardless of the height ASL, at which the measurements are made.

In between 43.5–43.1°, lidar SWS increases as a function of the distance from the coastline. Such increase is not unexpected and is due to the acceleration of the flow associated with the land/sea roughness transition. However, it is further amplified by the presence of the sheltered region (characterized by weaker winds) in the lee of the Massif Central (Figure 4b). In other terms, in the absence of such a feature, lidar SWS increase might not have been pronounced. This behavior is in good agreement with the ALADIN SWS obtained from ALADIN at 1800 UTC.

In between 43.2 and 42.8° (the core of the perturbed Mistral flow [Flamant, 2003]), the lidar SWS was relatively constant (12–13 m s⁻¹). The agreement with ASIS SWS (at 1600 and 1700 UTC) and TOPEX SWS was good. ALADIN SWS at 1800 UTC also were in excellent agreement with observations. At 43.1°, lidar SWS and ALADIN SWS are in good agreement with Atlante SWS at 1700 UTC. The Atlante SWS at 1600 UTC was 2.5 m s⁻¹ lower than at 1700 UTC. Such a trend as also been observed for the ASIS SWS, though not as marked (i.e., a 1 m s⁻¹ increase was observed between 1600 and 1700 UTC). It is thought to be related to the nonstationary aspect of the Mistral flow over the Gol, as well as the important spatial variability in the wind field due to the presence of a drifting sheltered region and low-level jets. ASIS and Atlante SWS measurements both showed a minimum in wind speed at approximately 1600 UTC, corresponding to the perturbed Mistral period [Flamant, 2003].

In between 42.8–42.5°, lidar SWS were observed to decrease by as much as 3–4 m s⁻¹, before increasing back to values around 15 m s⁻¹ (in connection with the presence of swell), these latter values being in agreement with TOPEX SWS retrievals. Note that this region of weaker winds was also observed in the ARAT SWS, so that it is not believed to be due to errors in the correction of atmospheric effects (changes in aerosol optical properties for example) in the lidar SWS retrieval procedure. Interestingly, ALADIN SWS also decreased substantially between 42.8–42.5°. However, ALADIN SWS kept on decreasing south of 42.5°. This decrease was interpreted by Flamant (2003) as the signature of a sheltered region in the lee of the Maritime Alps resulting from the interaction of the northeasterly synoptic winds and the obstacle. ALADIN SWS at 2100 UTC also showed that this feature had moved eastward as the Mistral became well established again. Hence, the picture arises that ALADIN might not have captured all the spatial variability associated with features such as sheltered regions.

Even though substantially larger, ARAT SWSs (at 100 and 300 m ASL) exhibited a trend similar to lidar SWSs, i.e., steadily increasing winds with fetch, a region of lighter winds and stronger winds again. The region of decreasing winds however was located slightly to the north of that observed in the near surface winds. This is consistent with the analysis of Flamant (2003) who showed that the flow at 300 m was more perturbed by the outflow coming...
from the Ligurian sea than at the surface. Hence, the location of the sheltered region was shifted to the northwest by the stronger outflow at 300 m.

TOPEX SWs were acquired later than any of the other measurements, during a period where the Mistral was no longer perturbed by the outflow from the Ligurian Sea as the cyclone was moving southward. The steadily increasing winds measured by TOPEX most likely revealed that the Mistral was well established again at the time of overpass.

Flamant (2003) has shown that the 1800 UTC ALADIN forecast compared well with the measurements made between 1600 and 1700 UTC. Therefore, even though TOPEX measurements were made at a time close to the ALADIN forecast shown in Figure 8, it is not so surprising to find that these retrievals exhibited the largest differences. This is confirmed by good agreement between TOPEX SWs and ALADIN SWs at 2100 UTC. Hence the TOPEX overpass occurred at a time when the Mistral was well established again.

4.5. Evolution of Roughness Length With Fetch

Figure 9 shows the temporal evolution of neutral SWs at 10 m ASL, significant wave height and wave phase velocity measured by ASIS between 1200 and 2400 UTC on 24 March 1998. In order to compute the roughness length in such nonstationary meteorological conditions, an attempt was made to correct for the time shift between the lidar and the radar measurements. This correction consisted of 2 steps: (1) because airborne measurements are not made instantaneously on leg AF, a correction was applied for the temporal drift of the variables between the beginning and the end of the leg (40 min and 47 min for LEANDRE 2 and RESSAC, respectively). The drift was computed from ASIS measurements, assuming the temporal evolution observed at this point was representative of that over the entire leg. The drift is null at the time of the aircraft overflight of ASIS (1632 and 1746 UTC for the ARAT and the Merlin, respectively); (2) because airborne measurements are not made simultaneously, the radar-derived Cp and Hs were shifted back 84 min, i.e., from 1746 to 1632 UTC, by increasing Cp by 1.42 m s⁻¹ and Hs by 0.05 m.

The evolution of lidar/radar-derived roughness length, z₀, as a function of the distance from the coast on leg AF is shown in Figure 10. For comparison, the roughness length derived from ASIS measurements between 1603 and 1733 UTC is also shown. z₀ exhibited the same trend as MSS and SWs: first, an increase with the distance from the coast, then a sharp decrease followed by a relatively constant value and finally an increase again with the distance from the coast. Between the coast line and 43.25°N, the increase of z₀ is believed to be caused by the sheltering effects in the lee of the Massif Central, and the fact that the lidar documented the transition between sheltered and unsheltered Mistral flow. Between 43.25°N and 43.0°N, we observed an increase of z₀ with the distance from the coast line (Figure 10). In this region the wave steepness
The spatial evolution of the drag coefficient. Given the sensitivity of \( C_{\text{sw}} \) to remotely sensed variables and the rather large uncertainties (10–25%) associated with these variables, we feel it is illusory to attempt to demonstrate that the drag coefficient can be unambiguously determined from airborne measurements alone.

For comparison, we show the drag coefficients estimated from ASIS measurements between 1603 and 1733 UTC in Figure 11. Note that the inverse wave age corresponding to these points is 0.0468, 0.0545, 0.0616. We also show, the inverse wave age dependent neutral drag versus wind speed relationships. As in the work of Draon et al. (2003), these relationships are obtained by solving iteratively a set of 3 equations, i.e.,
Figure 9. Temporal evolution of (a) neutral SWS at 10 m ASL, (b) significant wave height and (c) wave phase velocity measured by ASIS between 1200 and 2400 UTC on 24 March 1998. For panel (a), the vertical solid line indicates the time of ARAT overflight of ASIS (1632 UTC). The vertical dashed lines on both sides indicate the start and end time of leg AF flown by the ARAT. For panels (b) and (c), the vertical solid line indicates the time of Merlin overflight of ASIS (1746 UTC). The vertical dashed lines on both sides indicate the start and end time of leg AF flown by the Merlin.

The drag coefficients estimated from combined radar/lidar measurements are observed to correspond to inverse wave age values between 0.05 and 0.07, which is remarkably consistent with the ASIS measurements. For further analysis of the evolution of the drag coefficient with fetch, we have plotted with different colors and symbols the data grouped in the 4 latitudinal ensembles determined from the behavior of the SRL and SWS as a function of fetch: between the coast and 43.25°N (sheltered region, group 1), between 43.25°N and 43°N (developing sea, group 2), between 43°N and 42.7°N (developed sea, group 3) and beyond 42.7°N (swell contamination, group 4). The lidar/radar derived data indicated that younger (older) waves were indeed observed at short (long) fetch. Data from groups 2 and 4 were identified to exhibit less scatter than data from group 1 and 3. The reason for this is believed to be that in the corresponding regions, the action of wind on the sea surface was more steady, less perturbed than in the sheltered region (group 1) or at the eastern edge of the Mistral (group 3) where forcing of the wave field is expected to be more transient than in established flow regions. However, the drag coefficient data from groups 2–4, exhibit steeper trends with $U_{10}$ than those associated with the constant-inverse-wave-age relationships.

4.7. Friction Velocity

In Figure 12, we show the spatial evolution of the friction velocity derived from combined lidar/radar measurements and forecasted by ALADIN (1800 UTC) as a function of latitude. Also shown are the friction velocity derived from ARAT in situ measurements made at 100 m ASL using a eddy-correlation technique between waypoints F and B, as well as the ASIS measurements at 1603, 1634 and 1702 UTC.

(a) The ALADIN and lidar/radar friction velocities at the location of ASIS were found in good agreement with the ASIS measurement at 1634 UTC. Good agreement is also found between in situ aircraft measurements and lidar/radar.
height and (c) wave. For panel (a), the vertical dashed lines from group 1 and 3. The vertical dashed lines in (b) and (c), the sheltered region (group 1) (group 3) where forcing to be more transient than in the drag coefficient data trends with $U_{10\text{m}}$ than those relationships.

Figure 9. (continued)

The present study shows that in the coastal environment, and under strong wind, cold-air outbreak type conditions, accurate lidar-derived SWS can be obtained provided that exogenous information on the stability conditions and the whitecap coverage dependence on SWS is available. Stability was shown to be crucial as it could introduce a large bias in SWS retrievals ($6.7$ m s$^{-1}$ on average on leg AF). Fortunately, this variable can be obtained, with a reasonable accuracy, from numerical weather prediction models. In limited fetch conditions, the knowledge of the whitecap coverage dependence on SWS was found not to be as critical (as that of stability). The impact (or the lack thereof) of the whitecap coverage dependence with SWS on lidar-derived SWS was assessed by comparing results obtained using (1) the Monahan [1986] relationship (for fully developed sea conditions) and (2) a relationship determined specifically on 24 March 1998. The knowledge of such a relationship may not be considered a crucial part of laser remote-sensing oriented experiments in coastal regions. Particular attention was given to lidar calibration and inversion approaches to ensure relevant, trustworthy SWS retrieval by lidar. The somewhat tedious processing that needed to be implemented was a direct consequence of the fact that the lidar used in this study was not developed for the primary purpose of measuring SWS. A lidar system specifically developed for the purpose of measuring SWS [e.g., a system with longer wavelength and multiple look angles] would alleviate some of the limitations of the calibration and inversion approach described in this paper.

The lidar-derived SWS evolution with fetch observed on 24 March 1998 in the afternoon was validated against in situ and remote sensing measurements made from a buoy and a ship as well as from the spaceborne altimeter TOPEX. The spatial variability in SWS observed with the airborne lidar was controlled by the structure of the wake regions downstream of the Massif Central and the Maritime Alps,
defining the longitudinal extension of the Mistral, as well as influenced by swell resulting from the action of a steady northeasterly flow coming from the Ligurian Sea in connection with intense Alpine lee cyclogenesis. These findings were supported by the other measurements.

[67] Using a formulation of the dimensionless roughness dependence with inverse wave age given by Drennan et al. [2003] (adapted to this case study), we then show that airborne lidar and radar measurements can be combined to provide insight into the evolution with fetch of roughness length, neutral drag coefficient and friction velocity. The remotely sensed variables used as input are the significant wave height and the wave phase velocity (radar) as well as SWS (lidar). Given the sensitivity of the surface roughness determined from the Drennan et al. relationship and the uncertainties associated with above mentioned lidar and radar variables, we felt it was elusive to attempt to determine the “absolute values” of roughness length, neutral drag coefficient and friction velocity from airborne measurements alone. Rather, we have used ASIS measurements to constrain the lidar/radar retrievals at the location of the buoy and focused on the analysis of the evolution of roughness length, neutral drag coefficient and friction velocity with fetch.

[68] Four distinct sea state regimes over a distance of 100 km could be identified from the remotely sensed variables obtained with this novel approach, in connection with atmospheric forcing (namely, the wake region downstream of the Massif Central, the established Mistral, the Mistral flow perturbed by the wake downstream of the Maritime Alps, the northeasterly flow from the Ligurian Sea). The dependence of lidar/radar derived drag coefficient with lidar-derived SWS for the 4 regimes was found to be remarkably consistent with the relationship derived from the buoy measurements. Finally, lidar/radar derived friction velocities were found in good agreement with the buoy and in situ aircraft measurements.

[69] Over the open ocean (neutral stability conditions) in moderate wind conditions (less than 10 m s⁻¹), Menzies et al. [1998] were able to retrieve SWSs, with a good accuracy and without a need for exogenous data (submarine reflectance, whitecap coverage, atmospheric stability conditions), using lidar measurements made in the framework of the Lidar In-space Technology Experiment which took place in September 1994. Lidar observations made at the global scale in the framework of future spaceborne missions such as GLAS (Geoscience Laser Altimeter System) and CALIPSO (Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observation) could be used to improve the analysis of turbulent heat fluxes in coastal regions when combined with radar and in situ data. The perspective of deriving SWS from lidar measurements in coastal regions where
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Figure 10. Roughness length evolution with fetch along leg AF. Crosses represent the roughness length estimated from ASIS measurements between 1603 and 1733 UTC.

Appendix A: Solar Radiation and Lidar Background Correction Procedure

In coastal waters, submarine reflectance can be enhanced due to the presence of Chlorophyll, inorganic suspended particles from terrestrial origin, yellow substances (in the Rhone river plume) or even mineral dust (deposited at the surface during Saharan dust outbreaks). As a result, submarine reflectance at selected wavelength can exhibit an important spatial variability at the scale of the GoL. Figure 13 shows the normalized water-leaving radiance, $nLw$, at 670 nm observed by the Sea-viewing Wide Field-of-view Sensor (SeaWIFS) pass over the GoL at 1138 UTC on 24 March 1998 (obtained from the SeaDAS software). The normalized water leaving radiance is approximately the radiance which would exit the ocean in the absence of the atmosphere and with the Sun at zenith. The normalized water leaving radiance is linked to the submarine surface reflectance $PSSI'$ by $PSSI' = \frac{TrLw}{F_a}$ where $F_a$ is the solar irradiance at the top of the atmosphere. In Figure 14, we show the correlation existing between the lidar system detected background light (at 732 nm) scattered in the direction of the telescope and the SeaWIFS normalized water leaving radiance at 670 nm extracted from Figure 13 along leg AF. The background light is the ambient atmospheric light scattered in the direction of the telescope acquired prior to firing the laser, which is in turn subtracted from the subsequent lidar return signals. A background light measurement is performed prior to the acquisition of each lidar signal profile. A correlation coefficient of 0.71 (explaining 53% of the observed variance) is obtained. A large part of the variance not explained by the correlation is likely to be related to solar-induced Chlorophyll fluorescence at 670 nm. Finally, some of the scatter observed in Figure 14 can also be explained by the fact that SeaWIFS and lidar measurements were not acquired at the same time. Given the high correlation coefficient, we are confident that solar radiation diffused by the suspended particulates and/or foam in the direction of the lidar telescope is indeed removed through "background" light corrections applied to the lidar data, even though the data presented here only partly supports those conclusions (due to the fact that observations were not made at the same time and wavelength).

Appendix B: Lidar Inversion and Lidar-Derived Sea Surface Reflectance

Sea surface reflectance is inferred from lidar measurements using the method described by Flamant et al. [1998], as

$$\rho = \frac{1}{\pi} \Delta e(\lambda, \theta, z) \frac{S(\lambda, z)}{S(\lambda, z_o)}$$

(31)
Figure 11. Drag coefficient versus wind speed, both 10 m ASL neutral, for leg AF data. The solid lines are the inverse wave age dependent drag versus wind speed relationships resulting from equations (16), (17) and (18). Drag coefficients are grouped in 4 latitudinal ensembles: between the coast and 43.25°N (blue triangles), between 43.25°N and 43°N (red diamonds), between 43°N and 42.7°N (orange asterisks) and beyond 42.7°N (yellow crosses). Crosses represent the drag coefficients estimated from ASIS measurements between 1603 and 1733 UTC. Note that the inverse wave age corresponding to these points are 0.0468, 0.0545, 0.0616 and 0.0754.

which accounts for the lidar signal S(λ, z₀) in the layer close to the surface, the lidar surface return signal S₀(λ, z₀) as well as the average backscatter coefficient K(λ, τ, z₀) measured in a layer of depth Δz, at an altitude z₀ + Δz above the surface.

[3] The largest uncertainty on lidar-derived SSR is due to the uncertainty on K(λ, τ, z₀) which is obtained from the well-known “lidar inversion” procedure (e.g., Flamant et al., 1998). In this procedure (see below), accurate (K(λ, τ, z₀) and hence SSR, estimates can only be obtained if the attenuation of the laser beam by aerosol particles is properly corrected for between the aircraft and the sea surface. This implies that the aerosol population in the atmospheric column sampled by lidar is known (this information enables to calculate a particulate backscatter-to-extinction ratio, BER, profile using the Mie theory). Furthermore, the inversion procedure used in this study also relies on an independent measurement of the scattering ratio (defined as the ratio of the total backscatter coefficient to the molecular backscatter coefficient) at some height in the region of the atmosphere sampled by the laser beam.

[4] During FETCH, the reference scattering ratio necessary to initialize the inversion procedure was measured in situ by a nephelometer on-board the ARAT. To compute the BER, an idealized coastal aerosol population has been defined, based on previous studies undertaken in this region and ship-borne measurements. To compensate, in part, the lack of aerosol measurements, the sensitivity of SSR to the BER and to the reference scattering ratio has been analyzed.

B.1. Reference Scattering Ratio

[5] The reference scattering ratio is obtained from in situ scattering coefficient measurements made at 3.9 km ASL by a Meteorology Research, Inc. (MRI) integrating nephelometer model 1550B mounted on the ARAT. It measures the scattering by both gases and aerosols around 550 nm. No correction for the molecular contribution is applied in real time. The design of the instrument implies that only scattering by extinction, and not scattering by absorption, is measured (Heintzenberg and Charlson, 1996). The sampled air used in this instrument was heated to maintain a relative humidity below 60%. The sampling error on the
The solid lines equations, and are estimated from ASIS, corresponding to these measurements. The friction velocity evolution with fetch along leg AF. measured scattering coefficient is estimated to be of the order of 10% as specified by MRI. The average value of the scattering ratio at 3.9 km ASL was about 2 (not shown). The statistical variability associated with these measurements was observed to be on the order of 5%, peak to peak. The largest resulting quadratic error (computed as the square root of the errors squared) accounting for measurement uncertainties and statistical variability is 12%. Hence, the reference scattering ratio at 3.9 km ASL was equal to 2.0 ± 0.1 for the entire flight.

B.2. Aerosol Composition and Vertical Distribution

During a Mistral event, the coastal marine aerosol is a dynamic reservoir of particles originating from diverse marine and continental sources. Sea-salt, dust, water-soluble, particulate organic matter and soot can be important contributors to the coastal marine aerosol composition. During the FETCH experiment, the aerosol composition was analyzed from samples made with a 13-stage low cascade impactor installed on the front deck of the Research Vessel Atalante [Sellegri et al., 2001]. Special attention was paid to water-soluble and sea-salt aerosols. Total carbon and soot concentrations were not measured during FETCH. Size distributions, between 0.5 and 15 μm aerodynamic diameter, were obtained from an active scattering spectrometer probe mounted on the mast located on the front deck of the ship. However, these surface measurements are likely not to be representative of the aerosol composition above the MABL surface layer. On 24 March 1998, the air masses had traveled over the continent prior to being sampled by the aerosol instrumentation on the Atalante. Over warm land, the continental ABL was observed to be about 2 km deep [Flamant, 2003]. Due to convection and turbulence, aerosols are generally observed to be well mixed in the vertical, over the depth of the continental ABL, forming a so-called pollution plume. When continental aerosol plumes are advected across the coastline they are forced to rise above the MABL [e.g., Flamant, 2003]. As a consequence, a considerable part of the continental ABL particles may be transported over thousands of kilometers without removal by washout processes. Numerous measurements have shown this over the coastal Atlantic Ocean for example [Flamant et al., 2000; Assmann et al., 2001]. Hence,
(partial) knowledge of the aerosol composition near the surface is not enough for our purpose. However, no measurements of the aerosol composition were made above the sea surface (i.e., from airborne platforms). Therefore, we have assumed the aerosol population over the GoL to be a mixture of the maritime and continental aerosol described by Ackerman (1998). The errors and uncertainties on SSR retrievals associated with this assumption are assessed through sensitivity analyses as detailed in the following.

B.3. Sensitivity, Errors, and Uncertainties on SSR Retrievals

Given the lack of aerosol distribution measurements above the MABL surface layer, we have considered the likely range of variation of the BER in a coastal region based on the work of Ackerman (1998). Ackerman (1998) computed the BER fluctuations with relative humidity (RH) for two aerosol (marine and a continental) models and at two wavelengths, i.e., 532 and 1064 nm. Aircraft measurements and ALADIN forecasts were used to determine the range of RH conditions encountered during the flight. RH ranged between 40 and 100% in the MABL, and between 15 and 40% ± 5% aloft. Considering Ackerman’s marine aerosol model, we find BER values in the MABL ranging between 0.033 and 0.04 sr⁻¹ at 532 nm, and ranging between 0.02 and 0.04 sr⁻¹ at 1064 nm. Considering Ackerman’s continental aerosol model, we find BER values between 0.022.
ait'borne platforms). Therefore, we used the aerosol population over the GoL to be a combination of marine and continental aerosol described in Ackerman [1998] between 15 and 1064 nm. Aircraft measurements during a sounding over the GoL (section 2) to constrain the lidar inversion and infer BER values characteristic of each layer. This profile (Figure 15) was acquired in a region where the MABL was approximately 0.7 km deep and the continental ABL approximately 1 km deep (top at 1.7 km ASL). A series of 100 lidar shots acquired in the vicinity of the aircraft sounding (the aircraft sounding was performed at a later time) were used for the lidar inversion.

In order to carry out a meaningful comparison of lidar and nephelometer extinction profiles, lidar extinction profiles were retrieved at 732 nm and transposed to a wavelength of 550 nm assuming a range independent value of the ngström coefficient. We then iterated on the BER values in the different layers and on the ngström coefficient values until the lidar extinction profile converges within 10% (on average) of the nephelometer extinction profile.

The best agreement between the lidar-derived and nephelometer-derived extinction profiles was found for values of 0.02, 0.018 and 0.025 sr\(^{-1}\) in the MABL and continental ABL, respectively.

In the following, we have used the extinction coefficient profile obtained at 550 nm from the in situ nephelometer and a unique solution. The BER value of 0.018 sr\(^{-1}\) in the continental ABL is in good agreement with the characteristic value for continental aerosol proposed by Ackerman [1998]. On the other hand, the BER value derived in the MABL is much less than generally observed in pollution-free conditions, i.e., when the lidar signal is dominated by water soluble, particulate organic matter and sea-salt aerosol. This low value could be an indication of the rather polluted air masses advected over the GoL during the Mistral/tromontane event. Such pollution aerosols were indeed sampled on-board the Research Vessel Atalante [Sellegri et al., 2001].

Finally, the lidar-derived atmospheric reflectivity in the MABL was reported to increase significantly between endpoints A and F [Flamant, 2003]. He related the increase of lidar reflectivity observed east of 4.4°E on leg AF to an
increase in aerosol concentration and/or a change in aerosol optical properties downwind of the city of Marseille and the industrial petro-chemical complex of Berre/Fos (43.4°N, 4.5-4.6°E, Figure 1). Hence, the BER value obtained near endpoint F may only be representative of the air masses having traveled from the Marseille region. To the west of 4.4°E, air masses could be characterized by commonly assessed BER values of the order of 0.03–0.04 sr⁻¹, rather than the value of 0.02 sr⁻¹ near endpoint F. Thus, we have analyzed the sensitivity of SSR (derived from the selected series of 100 lidar shots) to the BER for a 1 km deep MABL in which the BER was varied between 0.02 and 0.04 sr⁻¹ (Figure 16). The sensitivity of SSR to the reference scattering ratio is also illustrated in Figure 16. Figure 16 shows that the SSR is very sensitive to the reference scattering ratio and only moderately sensitive to the BER. For a scattering ratio equal to 2, the SSR varied between 0.053 and 0.056 for the BER range considered, i.e., about 5%. For a BER of 0.035 sr⁻¹, the SSR varied between 0.05 and 0.057 for a scattering ratio ranging between 1.9 and 2.1, i.e., about 15%. This means that the SWS estimates are not sensitive to the evolution of the composition of the aerosol.

In this study, the lidar data along leg AF have been processed using BER values of 0.02, 0.018 and 0.025 sr⁻¹ in the MABL, continental ABL and free troposphere, respectively. The continental ABL top height was taken at 1.8 km ASL and remained unchanged along AF. The structural characteristics of the MABL along leg AF are derived from lidar measurements described by Flamant [2003]. These measurements [see Flamant, 2003, Figure 14c] showed the internal thermal MABL to reach a depth of 1.2 km at approximately 4.5°E. East of 4.5°E, the MABL structure characteristics over the sea changed dramatically: it was observed to be shallower, with a depth of about 0.7 km. The backscatter coefficient above the sea surface is then derived from each profile along AF and is used to compute the SSR using equation (B1). As mentioned earlier, a constant reference scattering ratio of 2.0 ± 0.1 at 3.9 km ASL was used for the entire leg AF. In the inversion procedure, the ±5% uncertainty on this parameter leads to a 10% uncertainty on the near surface backscatter coefficient [e.g., Flamant et al., 1998].

Appendix C: RESSAC and TOPEX Surface Mean Square Slope Measurements

The radar backscatter cross section (RBCS) per unit area (denoted σ₀, or Normalized Radar Cross Section NRCS) due to specular reflection on a rough surface has been shown by Barick [1968] and Hulsenzuel [1978] to be related to the probability density function (pdf) of ocean surface slope as:

\[ σ₀ = \rho^2 \frac{1}{\cos \theta} \left( \delta'_0 \delta'_0 \right). \]
where $\theta$ is the radar incidence angle and $\rho^2$ is an effective normal incidence reflectivity. The sea surface is defined by the height function $C_2$, for two orthogonal directions, and the slope $p$ is defined by $p = \sqrt{C_{21}^2 + C_{22}^2}$.

If we assume an isotropic rough surface with Gaussian statistics, equation (C1) becomes:

$$\cos \theta \cos \beta = \rho^2 \exp \left( -\frac{\tan \beta}{2} \right).$$

where $(\tan \beta)$ is the MSS of the slopes pdf.

To estimate the MSS values from the RBCS data, we analyze the dependence of the NCRS with incidence angle as described by Hauser et al. [1992]

$$\left( \frac{\tan \beta}{2} \right) = \frac{B}{\rho^2},$$

where $(\sin \beta)$ is in dB. The MSS values than RESSAC, we adjusted the Fresnel coefficient obtained in C-Band (0.61) by adding a wind of 7 to 15 m s$^{-1}$.

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