Aeronautical diagnostics for Clear-Air Turbulence forecast at Meteofrance in the context of DELICAT* european project.

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Abstract. A study on Clear-Air Turbulence (abbreviated by CAT) forecast in a Numerical Weather Model is presented in this paper. The main objective of this study is to evaluate ARPEGE Meteofrance-NWP model’s ability to reproduce CAT, by calculating various CAT indices at the regional scale (over Europe) in this model. The list of indices used here is inspired from that proposed by R. Sharman & Wolff (2006). Calculated indices are then compared with AMDARs (Aircraft Meteorological DAta Relay) turbulence measurements during winter, early in 2010.

This work was performed within DELICAT european project (*DEmonstration of LIdar based Clear-Air Turbulence detection), in the Seventh Research Framework program of the European Union [FP7], in Meteofrance national weather agency.

1. Introduction

Clear-Air Turbulence is revealed in shakings of in-flight aircraft due to small-scale atmospheric turbulent eddies in a clear atmosphere without convective or stratiform clouds, as defined in G.P. Ellrod & Ebernberger (2003). High altitude Clear-Air Turbulence remains a significant forecast problem for aviation in commercial roads, since serious damages can be caused suddenly and unexpectedly by this CAT phenomena. As air traffic has increased, more precise determination of CAT is required (R. Sharman & Wolff, 2006), to improve flight conditions and reduce additional costs (fuel losses (Ellrod & Knapp, 1991), flight delays, etc.). To prevent situations of CAT by more useful CAT forecastings and to assist pilots in avoiding CAT, it is necessary to better understand the various mechanisms of emission of CAT (Lane & Sharman, 2008). Generally, three known mechanisms can cause CAT: the Kelvin-Helmholtz instability (Howard, 1961; Miles, 1961), the anticyclonic flow (Knox, 1997), and the gravity wave breaking, as those generated by mountains (Jiang & Doyle, 2004; J.D. Doyle & Bartels, 2005; T.P. Lane & Watson, 2009), but these different types of turbulence are still not well constrained.
2. Statistics on AMDARs meteorological data

2.1. Meteorological AMDARs data

AMDAR is an international program with air carriers that provides automated weather observations, in the context of World Meteorological Organisation (WMO). EUMETNET-AMDAR is the same program reduced to some european partners in EUCOS Operational Program (2007-2012). In this program, integrated sensors on board commercial aircrafts measure atmospheric conditions encountered during the flight. These meteorological data are then transmitted back to the ground in real-time, using the meteorological network. AMDARs messages basically consists of two components: operational and positional data (time, altitude, geographical coordinates, etc.), and meteorological data (temperature, winds, turbulence). In E-AMDAR program, two parameters are relevant to turbulence calculated on-board: it is the “Turbulence Intensity” (IT), and the “Derived Equivalent Gust Velocity” (DEVG).

The Turbulence Intensity (in g) in AMDARs messages corresponds to the vertical acceleration measured by embedded accelerometers in the aircraft gravity center, and converted according to the following scale of WMO. Turbulence levels given by IT are: light (0.15 g, code 9), moderate (0.5 g, code 10), and severe (1. g, code 11). This parameter depends on aircraft characteristics.

The DEVG parameter (in m s$^{-1}$) developed by the Australian Defense Force (Sherman, 1985) gives the measure of the vertical gust velocity with respect to the vertical velocity in sea level. This parameter does not depend on aircraft characteristics. Below 2 m s$^{-1}$, no turbulence is considered, and above, three turbulence levels are usually defined from the range of 2 m s$^{-1}$ to 9 m s$^{-1}$ and more. In fact, some peaks of extreme turbulence can be found up to 30 m s$^{-1}$ and more. DEVG can be seen as a maximal instantaneous value of vertical velocity encountered by the aircraft.

AMDARs data have two particularities: inhomogeneous in space (preferentially concentrated around airports), and depending on air traffic (concentrated during the day time, especially in the morning).

2.2. Statistics on AMDARs data in early year 2010

Some preliminary statistical studies were made to determine the privileged period for CAT analysis with AMDARs data. The statistical sample used here corresponds to AMDARs data on EURAT01 domain (Europe: 3200-42E/20N-72N) at the beginning of the year 2010, for altitudes up to 7 kilometers high, with no turbulence induced by convection. Lightning observations were used to filter this sample of AMDARs data and to eliminate cases of turbulence-induced convection: if no lightning impact is detected around 50 km and +/- 30 min by Meteofrance or MetOffice measuring stations, data is kept.

In these conditions, the sample used here represents 0.3% of total AMDARs data, and 11% of valid AMDARs data of turbulence. Taking into account the test on the altitude, lightning cases corresponds to 2% of turbulent data, against 98%. Different classes of turbulence intensities were defined to take into account both informations on turbulence supplied by AMDARs data (cf. Figs. 1): light turbulence is considered if $IT = 9$ or $2 < DEVG < 4.5$; moderate if $IT = 10$ or $4.5 < DEVG < 9$; and severe if $IT = 11$ or $9 < DEVG$. In this sample of AMDARs data, 88% of turbulent cases represents light turbulence, against 8% for moderate turbulence, and only 4% for severe turbulence. Severe turbulent cases are more numerous in IT observations than in DEVG. Turbulent cases are concentrated in the first three months of the year 2010, which represents about ~75% of turbulent cases, indicating that the privileged period for CAT analysis is rather situated in winter, in the beginning of the year.
Figure 1. Statistics on AMDARs turbulent data, for the period from January to May 15\textsuperscript{th} of 2010, on EURAT01 domain (Europe), for altitudes up to 7 km high. On left: number of measured turbulence cases according to turbulence intensity without lightning (9: light, 10: moderate, 11: severe), and depending on months of early 2010 (red: January, yellow: February, turquoise: March, purple: April, pink: May). On right: number of measured turbulence cases represented in stacked areas, according to months and to lightning measuring stations (Meteofrance without lightning in blue, and MetOffice without lightning in yellow). Lighting cases were also represented (lightning cases measured by Meteofrance in red, and by MetOffice in turquoise).

By plotting AMDARs turbulent data on EURAT01 domain (cf. Figs. 2), various predilection areas of turbulence were found: many measurements of strong turbulence were detected over the Atlantic channel at the beginning of year 2010, and some isolated cases of extreme turbulence (greater than 9 m s\textsuperscript{-1}) were found over Central Europe, both in IT and DEVG.

Figure 2. Projection maps (latitude-longitude) of AMDARs turbulent data (DEVG or IT turbulent values) without lightning, up to 7 km high, during the first three months of year 2010, on EURAT01 domain (Europe). On left: for DEVG values (in m s\textsuperscript{-1}): no turbulence, light and moderate turbulence (∼ 0 – 9 m s\textsuperscript{-1} in blue), strong turbulence (∼ 9 – 14 m s\textsuperscript{-1} in turquoise blue), and extreme turbulence (up to ∼ 14 m s\textsuperscript{-1} from green to red). On right: for IT values (no units in WMO scale): light turbulence (in turquoise blue), moderate turbulence (in green), and strong turbulence (in red).
3. Aeronautical diagnostics with Meteofrance model

3.1. Description of Meteofrance model
To calculate aeronautical diagnostics for turbulence forecast, I made a simulation with the latest version (2010) of ARPEGE meteorological model of Meteofrance in order to take into account the latest developments of this operational model and, to use the latest assimilated data. This simulation experiment offers a global coverage on EURAT01 domain (Europe) with a mesh of 0.1° (approximately ~ 10 km). Isobars were chosen as vertical coordinates between 10^4 hPa (~ 100 m) and 150 hPa (~ 13.7 km). The vertical resolution of ARPEGE native model (cf. Fig. 3) was used in output, for altitude levels which interest us, between ~ 5.2 and 13.7 km. In fact, layer thickness is increasing with altitude between 0 m at the ground, and ~ 450 m at 13.7 km (~ 420 m at 10 km). The period of the run is the first three months of year 2010, corresponding to the period of AMDARs measured turbulence. Outputs were firstly made every hours.

![Variation of the vertical resolution of ARPEGE meteorological model](image)

3.2. Set of turbulence indices
Various conventional indices were calculated with ARPEGE meteorological model inspired by the set of turbulence indices used in R. Sharman & Wolff (2006), such as Richardson, Brown, Ellrod indices, Turbulent Kinetic Energy (TKE), Eddy Dissipation Rate (EDR), jet intensity, vertical wind shear, potential vorticity (PV), horizontal gradients of temperature and PV, Colson-Panofsky, and CATMOS indices (see Appendix for detailed formula).

TKE and EDR were directly implemented into ARPEGE model as new pronostic variables, while other parameters were calculated by post-processing. In fact, since the fourth meeting of METLINKSG in 1998, EDR parameter was designed as a standard parameter for embedded measure of turbulence, but a finer and more exhaustive validation is still awaited.

Overall, indices obtained with this simulation experiment have orders of magnitudes in good agreement with what can be found in literature, even if in details, for each indices, exact thresholds of levels of turbulence intensity and extrem values vary from one paper to another.

3.3. Case studies of strong turbulence
Case study was made in order to do a first comparaison between model outputs and AMDARs data. EDR parameter allows to visualize small turbulent structures, which evolve as “turbulent
cores” (cf. Fig. 4) with size range between ~ 40 and 100 km, generally elongated along the
direction of the wind, with a few kilometers thick vertical, and with a variable life-time range
between a few hours to one or two days approximately. These “turbulent cores” are structures
highly localized in space and time, making their detection more difficult in view of uncertainties
that arise in both the model and observations.

From AMDARs analysis, two cases of strong turbulence were selected for analysis of
turbulence-generating mechanisms in the model.

3.3.1. A case study of strong turbulence generated by the interaction between the jet and
mountains. The first case of observed turbulence corresponds in the model, to a turbulence
generated by the interaction between the jet and Turkey mountains (cf. Figs. 4), and which
probably results in mountain wave out-breaking above Carpathians mountains. In this turbulent
case, EDR and TKE calculated in ARPEGE model both present a local maximum corresponding
to a “turbulent core”, which is pushed by the south-east wind (the jet maximum value is ~
42 m s\(^{-1}\)) and which is moving around a big depression situated over Central Europ at that time.
The Richardson and Colson-Panofsky indices each show a positive local minimum (~ 0.7 for
Richardson), in a context of positive vertical gradient of potential temperature (~ 0.003 K m\(^{-1}\)),
indicating that the “turbulent core” is associated to a less stable area in a stable surrounding
environment. The vertical wind shear is very strong all along Carpathians mountains (the
maximum is ~ 0.02 s\(^{-1}\)). This “turbulent core” takes place in an area of sign changing of
potential vorticity, and the horizontal gradient of potential vorticity presents alternating positive
and negative bands, parallel to Carpathians mountains. This probably indicates the presence
of instabilities in the flow. Ellrod indices also presents strong local maxima (~ 6 \(10^{-6}\) s\(^{-2}\) for both
indices), as well as the horizontal gradient of temperature (~ 11 \(10^{-5}\) K m\(^{-1}\)), and CATMOS
indices (\(CATMOS_1 \sim 4 \times 10^{-5}\) m s\(^{-2}\) and \(CATMOS_2 \sim 8 \times 10^{-7}\) m s\(^{-2}\)).

In a vertical cross section along Carpathians mountains (cf. Figs. 4), the “turbulent core”
clearly appears on the equatorial side of the jet in the same altitude. The vertical wind shows
alternating upward and downward movements in this turbulence area, and oscillations were
found in the potential temperature vertical structure. All these clues indicate that atmospheric
waves were probably out-broken at this altitude near the tropopause.

3.3.2. A case study of strong jet-generated turbulence. Firstly, I have researched some turbulence
in the model over the Atlantic channel, where many AMDARs turbulence measurements
were detected. Some turbulence was modeled by ARPEGE Meteofrance model in this area, but
with a temporal uncertainty of the order of the day, and with very low intensities. However,
strong turbulence was modeled further upstream of the north-westward Atlantic jet: this tur-
bulence was generated in front of the jet and on its equatorial flank over the Atlantic ocean
(cf. Figs. 5), in good agreement with what can be found in literature, and leading to strong
intensities of TKE and EDR. We can clearly see in the model some “turbulent cores” pushed to
the front of the jet, and vanishing upon reaching the Atlantic channel. In this turbulent case,
strong upward movements are presents in front and on both sides of the jet, associated with
sursaturated areas (greater than 100%), probably indicating the presence of convection.

3.4. Temporal correlation between aeronautical diagnostics
A study on the temporal evolution between various calculated indices in the model have been
made with different domain size and timesteps, in order to evaluate if indices evolve in the same
way in the vicinity of a turbulent event, and if it presents some temporal correlation. The tur-
bulent case over Carpathians mountains have been chosen for this study. Firstly, indices have
been analysed over an enlarged domain size of 12° corresponding to a square of side ~ 1200 km
Figure 4. Over Central Europe on January 11th of 2010 at 1 h between 250 – 300 hPa, with ARPEGE Meteofrance model. On top: Projection maps of EDR (on left, in m² s⁻³), and EDR (in blue contours) plotted on jet intensity (on right, in m s⁻¹). On bottom: Vertical cross sections (vertical pressure in hPa versus a direction along Carpathians mountains) with Carpathians mountains in turquoise blue, of EDR (in blue contours) plotted on jet (on left), and plotted on potential temperature (on right, in K).

Figure 5. Over Atlantic Ocean on January of 2010 between 250 – 300 hPa, with ARPEGE Meteofrance model. On Top: Projection maps of EDR (in m² s⁻³, in blue contours) plotted on jet intensity (in m s⁻¹), at two different times: on the 15th at 12 h (on left) and on the 16th at 1 h (on right). On bottom: Vertical cross sections (vertical pressure in hPa versus a direction perpendicular to the jet) of EDR (in blue contours) plotted on jet (on left, in m s⁻¹), and plotted on relative humidity (on right, in %).

centered on Carpathians mountains, and I have looked their temporal evolution during four days centered on the observed and modelled turbulent event, with a timestep of one hour, between two model levels 250 and 300 hPa. Some temporal correlation were found in averaged TKE, EDR, Ellrod indices, Dutton and the horizontal gradient of temperature, as well as in maxima of TKE, EDR, and Ellrod indices (cf. Figs. 6). All of these indices show a peak with strong values at the time of the turbulent event, whereas Richardson number and Colson-panofsky indice present minimum values at that time.
Figure 6. Temporal correlations between calculated indices of CAT forecast in ARPEGE meteorological model, on a domain centered on Carpathians mountains of \( \sim 1200 \) km side, during four days (from January 9th at 0 h to January 13th at 0 h of 2010), with a timestep of one hour, between 250 – 300 hPa. Spatially averaged fields of EDR and TKE (on top, in ISU), and Ellrod indices, Dutton and the horizontal gradient of temperature (on the middle, in ISU). Minimum fields with the same conditions as above for Richardson and Colson-Panofsky indice (on bottom, in ISU). Warning: all indices were rescaled to be plotted on the same figure.

By varying the size of the domain analysis from \( \sim 1200 \) km to \( \sim 250 \) km, we have always found the peak value corresponding to the turbulent event more or less pronounced. If we focus well on the turbulent event, we obtain a peak even more pronounced than the domain size is small, both in averaged and maxima indices, but the risk is to miss the turbulent event.

By varying the timestep from 1 h to 10 min, there is really no new informations brought by a more finer temporal resolution in this case study.

4. Summary

This work shows the relevance of a joint analysis of various indices for turbulence forecast in a meteorological model.

The two turbulent case studies have different physical origins in ARPEGE meteorological model, resulting in two different combinations of indices, each with different weights, highlighting the interest of R. Sharman & Wolff (2006)’s study.

In this model, the Eddy Dissipation Rate (EDR) seems to be a very convenient field to visualize the evolution of these transient and spatially-localized turbulent structures, that are advected by the atmospheric flow. Other diagnostics for turbulence forecast are very useful to better target physical conditions that have led to this turbulence. These are not absolute
intensities of fields that are relevant for turbulence forecast, but rather the local gradient versus the surrounding environment. In fact, in the case of Carpathians mountains’ turbulence, for example, the EDR maximum intensity corresponding to “turbulent cores”, is rather low compared to intensities which are usually found in the literature for strong turbulence (maxima are $EDR \sim 0.002 \, m^2 \, s^{-3}$ and $TKE \sim 2.9 \, m^2 \, s^{-2}$), but the deviation from the ambient environment is locally strong. These strong local gradient present in most indices of turbulence, allow us to visualize and detect these “turbulent cores”, clearly coming out from the surrounding environment.

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Appendix

Formula used in this study to calculate aeronautical diagnostics of Clear-Air Turbulence forecast are summarized in this section. This list of turbulence diagnostics is largely inspired from that used by R. Sharman & Wolff (2006).

- The Richardson number (without units) gives a local measurement of atmospheric stability.

$$R_i = \frac{N^2}{S_v^2}$$  \hspace{1cm} (1)

with

$$N = \sqrt{\frac{g}{\theta}} \frac{\partial \theta}{\partial z}$$  \hspace{1cm} (2)

being the Brunt-Vaisala frequency (in s$^{-2}$), $z$ the altitude (in m), $g = 9.81 \, m \, s^{-2}$, $\theta = T(P_0/R_a/C_p)$ the potential temperature (in K), with $P_0$ the standard atmospheric pressure, $R_a$ the perfect gas constant of air, $C_p$ the specific heat capacity at a constant pressure, and

$$S_v = \left| \frac{\partial V}{\partial z} \right| = \sqrt{\left( \frac{\partial U}{\partial z} \right)^2 + \left( \frac{\partial V}{\partial z} \right)^2}$$  \hspace{1cm} (3)

the vertical shear of atmospheric flow.

$U$, $V$, $W$ are respectively horizontal and vertical components of the wind (in m s$^{-1}$).
Ellrod indices (in $s^{-2}$) are derived from simplifications of frontogenesis function (Ellrod & Knapp, 1991), most important elements being the intensity of horizontal gradient of potential temperature (proportional to the vertical wind shear with the thermal wind equation), and the deformation of the flow.

\[ TI_1 = DEF S_v \]  
\[ \text{with } S_v \text{ the vertical wind shear defined in equation 3, and } DEF \text{ the horizontal deformation of the wind (in } s^{-2} \text{), calculated with:} \]

\[
\begin{cases} 
D_{st} = \frac{\partial U}{\partial x} - \frac{\partial V}{\partial y} \\
D_{sh} = \frac{\partial V}{\partial x} - \frac{\partial U}{\partial y}
\end{cases}
\]

$D_{st}$ being the horizontal deformation of the wind due to horizontal stretching, and $D_{sh}$ the deformation due to horizontal wind shear. The resulting deformation is:

\[ DEF = \sqrt{D_{st}^2 + D_{sh}^2} \]

A convergence term were added to the first Ellrod indice to define the second Ellrod indice as follows:

\[ TI_2 = (DEF + C_{vg}) S_v \]

with $C_{vg}$ (in $s^{-1}$), the horizontal convergence:

\[ C_{vg} = -\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} \]

Dutton indice (in $kg \ m^{-1} \ s^{-3}$) was defined by Dutton (1980) from linear regressions in analysis of turbulence reports over North Atlantic and Ouest europe in year 1976.

\[ Dn = 1.25 \ S_h + 0.25 \ S_v^2 + 10.5 \]

$S_v$ is the vertical wind shear defined in equation 3, and

\[ S_h = \frac{1}{|V|^2} (UV \frac{\partial U}{\partial x} - U^2 \frac{\partial U}{\partial y} + V^2 \frac{\partial V}{\partial x} - UV \frac{\partial V}{\partial y}) \]

is the horizontal wind shear.

Jet intensity (in $m \ s^{-1}$), as well as the vertical wind shear were connected to turbulence. In fact, Endlich (1964) noticed that turbulence linked to the jet stream was stronger when the jet intensity or/and vertical shear was/were stronger.

\[ |V| = \sqrt{U^2 + V^2} \]

Horizontal gradient of temperature (in $K \ m^{-1}$) were used by numerous forecasters for airlines, and by G.S. Buldovskii & Rubinshtejn (1976). It is linked to the measure of horizontal deformation, as well as vertical wind shear (according to the thermal wind equation):
\[ |\nabla_h T| = \sqrt{\left(\frac{\partial T}{\partial x}\right)^2 + \left(\frac{\partial T}{\partial y}\right)^2} \] (12)

- The Turbulent kinetic energy (in \(m^2\ s^{-2}\)) is calculated on the basis of the equation of turbulent kinetic energy. It is defined by:

\[ TKE = (U'^2 + V'^2 + W'^2) \] (13)

where \(U', V',\) and \(W'\) are wind field fluctuations against the mean flow.

- The dissipation of the turbulent kinetic energy (in \(m^2\ s^{-3}\)) is given by:

Following the recommendations of the fourth meeting of METLINKSG (METeorological information data LINK Study Group) in the International Civil Aviation Organization (OACI), appendix number three was modified to adopt \(EDR\) as the most suited parameter for embedded measure of turbulence.

\[ EDR = \frac{C_T TKE^{3/2}}{\ell_c} \] (14)

where \(0.7 < C_T < 0.85, \ell_c\) is a characteristic length of turbulence dissipation, calculated by Bougeault & Lacarrre (2005) in ARPEGE meteorological model in Meteofrance.

- Colson-Panofsky indice (in \(m\ s^{-2}\)) was established by D. Colson (1965) on dimensional considerations in a stable atmosphere. This indice allows to estimate the intensity of turbulence in a clear sky:

\[ I_{cp} = \frac{TKE_{cp}}{\ell_c^2} = \left| \frac{\partial V}{\partial z} \right|^2 (1 - \frac{R_i}{R_{ic}}) \] (15)

where \(R_i\) is defined in equation 1. \(R_{ic} = 0.4\) is a critical Richardson number.

- Potential vorticity (Knox, 2011, in \(K\ m^{-2} s^{-1} kg^{-1}\)) and its horizontal gradient (Shapiro, 1978, in \(K\ m s^{-1} kg^{-1}\)):

\[ |PV| = | - g \xi_a \frac{\partial \theta}{\partial P} | \] (16)

\[ |\nabla_h PV| = \sqrt{\left(\frac{\partial PV}{\partial x}\right)^2 + \left(\frac{\partial PV}{\partial y}\right)^2} \] (17)

with \(P\) (in \(Pa\)) the atmospheric pressure, \(\xi_a\) the absolute vertical vorticity (in \(s^{-1}\)).

\[ \xi_a = \xi + f = \frac{\partial V}{\partial x} - \frac{\partial U}{\partial y} + f \] (18)

where \(\xi\) is the relative vorticity of the wind. \(f = 2\Omega \sin(\phi)\) is the Coriolis parameter (in \(s^{-1}\)), with \(\Omega = 2\pi/R\) the planetary rotation, \(R\), its radius, and \(\phi\), the latitude.
• Some indices related to wind intensity (for CATMOS$_1$) or vertical gradient of temperature (for CATMOS$_2$) and the horizontal deformation of wind were also developed to improve turbulence forecasts. Following diagnostics (in m s$^{-2}$) have been revealed very useful in some statistical studies for CAT prediction (Reap, 1996).

\[ CATMOS_1 = |V| \text{DEF} \quad (19) \]
where $|V|$ were defined in equation 11 and DEF in equation 6.

• Correlation entre le gradient vertical de température et la déformation (in K m$^{-1}$ s$^{-1}$):

\[ CATMOS_2 = |\frac{\partial T}{\partial z}| \text{DEF} \quad (20) \]

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