In-Situ Measurements of Temperature Fluctuation and Vertical Wind Shear for Altitudes up to 30 km in the Atmosphere

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Abstract

An atmospheric profiling system has been developed at the University of Tennessee Space Institute. The system consists of temperature, pressure, humidity, relative wind, and a Global Positioning System GPS to track the payload's position. A data logger unit was integrated into the instrument for saving measured atmospheric properties. The atmospheric profiling system was carried through the troposphere and up to the middle region of the stratosphere layer using high altitude balloons. The balloon is assumed to follow a Lagrangian path where the relative horizontal wind represents turbulent perturbations (i.e. gusts). The Parcel theory (a numerical model) used in meteorology to predict atmospheric instability and the Richardson number schemes used for turbulence were adopted in investigating the presence of turbulence indicators. Theories and terminologies relevant for the analysis of turbulence are described and the results from both methods were analyzed and compared.

Table of Contents

List of Tables

List of Figures

Chapter 1: Introduction

1.1 Overview

The highest-flying research aircraft is the U-2 derived NASA ER-2 with a ceiling of 70,000 feet (21.3 km). In-situ measurements above this altitude are restricted to balloon and sounding rockets. As a result, information on turbulence scales above ~21km is sparse. Here we are measuring the in-situ turbulence from the surface to 105,000 feet (32 km) primarily utilizing wind velocity and meteorological measurements on a balloon payload.

Turbulence is assumed to exist solely in the turbosphere (surface to 80km), above the turbosphere is assumed to be purely laminar flow with mixing by diffusion only. Our main atmospheric layers of interest are the troposphere, jet stream, tropopause and lower stratosphere. The troposphere is defined as the lowest region (surface to 11km) where temperature decreases with height. Colder air directly over warmer air is unstable at the smallest vertical scale. At the larger scale, this local instability is bounded by the adiabatic heating of the air parcels with compression, and the adiabatic cooling of the air parcels with expansion (First law of thermodynamics). This combination of instability at the small vertical scale and stability at the large scale give us our "weather".

The stratosphere (above ~15km) warms with height creating stability at both the small and large vertical scales. In between the troposphere and stratosphere lies the tropopause, which is commonly assumed to be isothermal, having neutral stability at the small vertical scale, and stability at the large scale. Complicating this system is the jet stream. Above Tennessee is the Northern Hemisphere polar jet (10 to 15 km), this Jet is formed at the interface of the Polar and Ferrel circulation cells and the Coriolis force acting on those moving air-masses. Commercial air traffic normally is influence only by the lower altitudes of the jet stream (up to 40,000 feet), a region known to pilots and passengers as the commonly experienced Clear-air turbulence. Routine air traffic does not reach altitudes associated with the upper altitudes of the jet stream – and therefore less is known of the turbulent properties at the upper edge of the jet stream (60,000 feet).

1.2 Background and Broader Impact

Meteorological data within the stratosphere layer is very limited. The amount of existing wind gust data and temperature perturbation observations are largely limited to altitudes below 25 Km; In-situ wind fluctuation data at higher altitudes are virtually nonexistent. The U-2 and ER-2 airplanes are able to measure atmospheric parameters for altitudes below 70,000 feet. In addition, space-bound or reentry air vehicles can record very limited data as it briefly transits the stratosphere. The National Oceanic and Atmospheric Agency (NOAA) has been successful in obtaining atmospheric in-situ measurement (using observational weather balloons). However, the data is limited to pressure, temperature, and humidity only. Several low altitude meteorological

UAVs, such as the Aerosonde, do exist but often-detailed information about the atmospheric sensor package is not published. Concerning the wind vector the only information available is that the Aerosonde is able to measure the wind, but it is not published how, and the measurement are limited to 30,000 feet above sea level.

Because of the lack of information concerning stability and turbulence within the stratospheric region, and due to the widely adopted assumption that this layer is considered stratified or stable; limited studies have been conducted regarding the existence of turbulence in stratosphere. Therefore, my objective was to develop an atmospheric profiling instrument that is capable of measuring and recording In-situ measurements of temperature, pressure, humidity, and relative wind turbulence/gusts for altitudes reaching 30 km. The recorded data was used in the analysis of turbulence, while two different methods were used in the investigation of stratospheric instability; the Parcel theory and the Richardson number scheme. The results and conclusions for this study are expected to lead to better understanding high altitude turbulence occurrences.

1.3 Intellectual merit

Improvement in computational atmospheric dynamics is exhibited by the capability of numerical weather models to define instability processes related to turbulence observed at altitudes above 25 km. As the mathematical weather simulation tools advanced, they can be used to cover estimates of atmospheric perturbations from the present gust database for airplane design at altitudes below 15 km to altitudes between 25 km and 50 km where hypersonic airplanes missions are conducted. The indeterminate potential for future air-breathing hypersonic flight research vehicles to encounter strong turbulence at higher altitudes could penalize the design of these vehicles by excessive cost or limitation and restrictions on performance. Since the atmospheric structure changes distinctly with altitude, direct extrapolation of gust magnitudes encounter probabilities from lower altitudes to the higher flight altitudes is not advisable.

The atmosphere is a complex dynamical system that cannot be predicted without the use of numerical models. These numerical weather prediction (NWP) models are computer codes based on mathematical representations of physical and thermodynamic processes. The results from these models offer information for daily weather forecasts as well as climate projections for the future decades. Hence, the fidelity of NWP simulations are of vital interest to the atmospheric science community. In-situ data collected here can be used to verify, validate, and even modify existing weather numerical models for accurate weather predictions and climate projections. Large data centers such as the U.S. National Weather Service (NWS) and the European National Centers for Environmental Prediction (NCEP), operationally use data surface based automatic weather stations, released radiosondes and satellite based sensors to monitor the state of the atmosphere. In addition to environment monitoring, the quality of data sets are crucial for the initialization of the NWS models and operational implementation In-situ measurements in data assimilation cycles [e.g. Hollingsworth et al., 1986 and Daley, 1991].

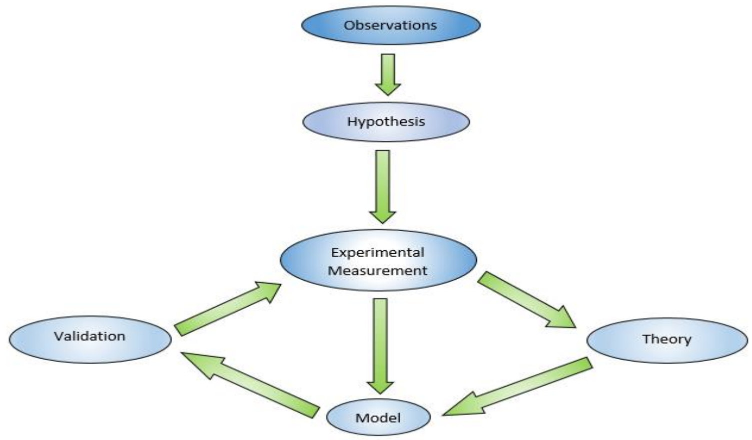

Figure 1: Sketch of the observed and theoretical cycle showing the equal importance of in-situ measurements and models

In general, the basic understanding of physical processes in daily weather originates from meteorological measurements. Observations offer the fundamentals for new hypotheses investigated in targeted meteorological field experiments, such as the investigation of turbulence within the stratosphere layer, which is the main objective of this research. Ideally, observational findings can be used to raise new theories, verify and improve already existing theories if necessary. The theoretical work finds its application in NWS models. To validate these models, measurements are always required because they represent the unique source of information on the “true” state of the atmosphere. However, one should be aware of the limitations, errors, and uncertainties associated with their measurement methods. Figure 1, outlines the continuous interaction between observations, model, and theory as tools. The use of measurements in the development and use of weather simulation models can be considered to be of equal importance in atmospheric sciences [Warner, 2011].

Chapter 2: Program Description

2.1 Atmospheric Thermal Structure

Originally, scientists thought that the atmospheric temperature decreased continuously with increasing altitude until it reaches absolute zero degree Kelvin. It was not until pioneering aerological experiments performed by *Abmann* and *Teisserenc de Bort* that led to the discovery of an unexpected phenomenon [*Assmann*, 1902; *Teisserenc* de *Bort*, 1902]. They observed a temperature increase above 10 km altitude and thus discovered the beginning of a new atmospheric layer, the stratosphere. In the following decades, the technical progress allowed more temperature sounding data measurements using weather balloons, sounding rockets, and airplanes (the ER-72 and the U2) enabled further research on the atmospheric characteristics and structure for altitude above 10 km. The sum of all these measurements resulted in a temperature profile which allows the division of the atmosphere into specific layers as shown in Figure 2.

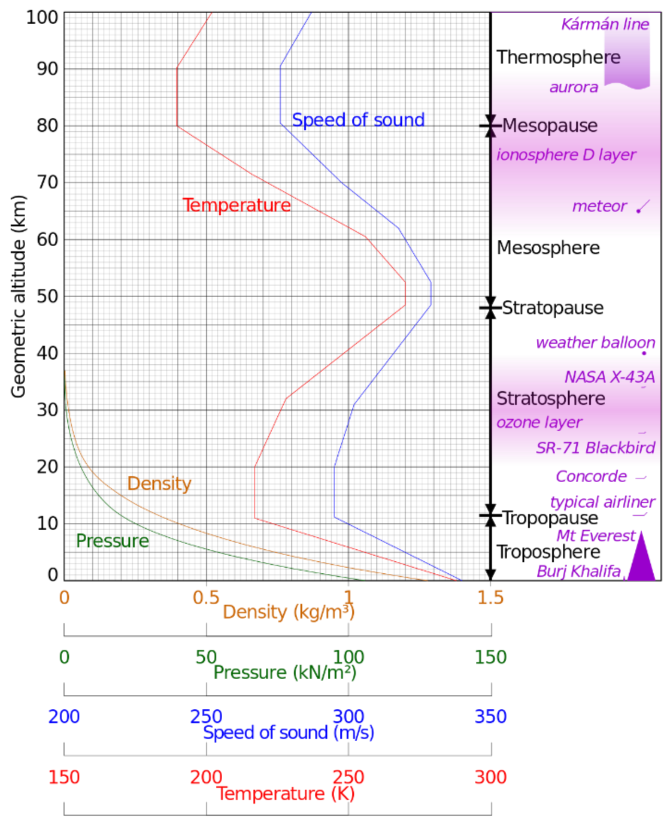

Figure 2: U.S. Standard Atmosphere graph of air temperature, pressure, density, and the speed of sound as a function of altitude

Figure 2 represent the temperature, pressure, and density profiles for a standard atmospheric day or conditions; that is when the temperature at sea level (zero altitude) is equal to 15 degree Celsius and the air pressure is 14.7 (*psi) or 101 (Kpa)* , however, the air density (ρ) is calculated using the ideal or perfect gas law equation, where T is temperature, P is the pressure, and R is the specific gas constant for dry air and is equal to 287 ($J.Kg^{-1}.K^{-1}$) or 1716 ($ft.lbf.slug^{-1}.R^{-1}$)

$$\rho = \frac{P}{R \times T} \tag{1}$$

Atmospheric layers are characterized by variation in temperature resulting from the absorption of solar radiation (visible light) at the surface, near-ultraviolet radiation in the middle atmosphere, and ultraviolet radiation in the upper atmosphere. The Troposphere is the lowest part (layer) of the atmosphere that extends to an altitude of ~ 10 km at the equator. Temperature and water vapor content in this layer decreases rapidly with increasing altitude; water vapor plays a major role in regulating air temperature through the absorption of solar energy and thermal radiation from earth's surface. All weather phenomena occur within the troposphere, although turbulence extends throughout the stratosphere. The upper boundary of this layer is known as the tropopause, which marks the transition between the troposphere and the stratosphere. The stratosphere is the second major strata of air in the atmosphere. It extends above the tropopause to an altitude of ~ 50 km above the surface. Air temperature within the stratosphere is relatively constant up to an altitude of ~ 20 km and then it increases gradually up to the stratopause. Because air temperature in the stratosphere increases with altitude, it does not cause convective buoyancy, and has a stabilizing effect on atmospheric conditions.

The ozone layer is contained within the stratosphere, which plays a major role in regulating the thermal regime of the stratosphere, as water vapor content within the layer is very low. Approximately 90% of the ozone in the atmosphere resides in the stratosphere. Solar energy within the ozone layer is converted to kinetic energy when ozone molecules absorb ultraviolet radiation, resulting in heating of the stratosphere; ozone absorbs the bulk of solar ultraviolet radiation in wavelengths from 290 nm to 320 nm (UV – B radiation). [*Labitzke and van Loon, 1999*]. Since this study focuses on measurements in the stratosphere, the characteristics of this layer and some important phenomena will be described in more detail below. The mesosphere extends up to the mesopause, which is approximately 85 km (depending on season). This layer resides above the stratosphere and it extends to altitudes of ~ 105 km depending on season, is characterized by decreasing temperature; the coldest temperature on earth is found on the top of this layer. In addition, a phenomenon like noctilucent clouds appear [*Gadsden and schoder, 1989*] also shown in figure 2. At the mesopause the air temperature stays relatively constant; above the mesopause the temperature starts to increase and the thermosphere layer of the atmosphere begins, this increase in temperature is due to the absorption of intense solar radiation (short wave) by the limited amount of oxygen molecules. At this extreme altitude, gas molecules are widely separated and air is said to be rarefied and it does no longer obey the ideal gas law.

Finally, the most distant atmospheric region from earth's surface is called the Exosphere. In this layer, fast travelling molecules can escape to space and slow moving molecules are pulled back to earth by gravity with small probability of colliding with another molecule. The exosphere is observable from space as the geocorona, seen to extend to at least 60000 miles from the surface. The exosphere is a transitional zone between earth's atmosphere and interplanetary space.

2.2 Stratosphere Layer Characteristics

As mentioned above, the temperature increase in the stratosphere is caused by the absorption of ultraviolet radiation by the ozone layer that resides in this layer. Brewer [Brewer, 1949] and Dobson [Dobson, 1956] have detected a global circulation within the stratosphere. This residual circulation consists of an upward air motion from the troposphere into the stratosphere within the tropics, a pole-ward transport within the stratosphere, and a downward motion in the middle and polar altitudes. Through this circulation, mass and trace gases are transported from the tropical tropopause to the extra-tropics [*Holton, 2004*]. The average time for an air parcel to be transported from the tropical tropopause to a given location in the stratosphere is defined as the mean age of stratospheric air [*Engel et al. 2009*]. So far, a wide discrepancy exists between the observations of the age of the stratosphere and numerical models used to predict it and further investigations are needed [Waugh and Hall, 2002; Waugh, 2009].
Scherhag [Scherhag, 1952] discovered another stratospheric phenomenon. A dramatic warming that occurs (within few days) during the winter season in the Northern Hemisphere (minor warming) can be accompanied by a total change of the stratospheric circulation (major warming) [*Labitzke, 1972*]. Although the mentioned effects influence not only the stratosphere but also the atmospheric layer above and below, the stratosphere has long been underestimated within climate models. Previously, the stratosphere has been considered as a kind of an upper border [*Labitzke and van loon, 1999*]. However, the role of the stratosphere has changed fundamentally due to experimental and theoretical studies over the past two decades [*Gerber et al, 2012*].

For instance, any long-term changes to the stratospheric winds and temperatures could possibly affect the surface's climate change [*Baldwin et al., 2007*]. Comprehensive climate simulations and experiments have revealed that the vertical coupling between the troposphere and the stratosphere can occur in both directions; changes in the stratospheric temperature or ozone concentration can lead to changes in the troposphere behavior. Consequently, a better representation of the stratosphere is needed to improve weather forecasts and climate predictions; therefore, further experiments and theoretical studies are needed in order to fully understand many stratospheric processes such as turbulence.

Due to its positive temperature gradient, the stratosphere is considered the atmospheric layer of stability and stratification; however, turbulence occurs due to breaking gravity waves and or strong wind shears, which induce Kelvin-Helmholtz instabilities. Stratospheric turbulence can appear on scales ranging from few millimeters up to several kilometers. Previous observations have shown turbulence occurs in thin isolated layers extending from ten to hundred meters vertically and to hundred kilometers in the horizontal direction [*Barat, 1982a; Sato and Woodman, 1982*]. Although turbulence and associated processes play an important role within many different aspects of the atmosphere, they have not been fully quantified or understood [*Wyngaard, 1992; Fritts at al., 2003*]. Furthermore, almost all numerical models dealing with atmospheric circulation, dynamics, energy, and composition must contain the effect of turbulence [*Gavrilov et al., 2005*]. Even though turbulence in the stratosphere is considered to be weak on average compared to mesospheric turbulence [*Lubken, 1992; Hcking, 1999*] it does affect a large number of atmospheric processes. For instance, the energy transfer from the troposphere to the mesosphere is modified by the energy dissipation within the stratosphere due to turbulence; therefore, further examinations of stratospheric turbulence is not only important for understanding the stratosphere itself, but for understanding the energy budget of the middle atmosphere as well.

In addition to the scientific interest, turbulence in the troposphere and lower stratosphere can be quite dangerous for aviation [*Sharman et al., 2012*]. An important parameter of turbulence for atmospheric modeling is the energy dissipation rate(ε), which represents the amount of energy dissipated or converted into heat. In the stratosphere dissipation occurs at the very small scales of only few centimeters or less; remote sensing such as radars, liadars, and satellite based sounders do not provide sufficient resolution to measure turbulence down to the meter scales, or provide any signal in the middle stratosphere [*e.g. Gurvich and Brekhovskikh, 2001; Luce et al., 2002; Engler et al., 2005; Samalikho et al., 2005; Sofieva et al., 2007*]. In-situ measurements are typically performed either below 23 km using aircraft [*e.g. Frehlich et al., 2003; Siebert et al., 2007*] or above 60 km using sounding rockets *[e.g. Lubken et al., 2002]*. In-situ measurements in the middle stratosphere are generally restricted to high altitude balloons. During the 1980s pioneer J. Barat and associates have used balloon-borne sonic anemometers for wind measurements *[Barat, 1982; Barat et al., 1984; Dalaudier et al., 1989],* the sounding data from their experiments resolved scales down to a few centimeters. The higher the resolution and precision of the measurements, the more accurate results are obtained for the energy dissipation rate and therefore a better estimate of the intensity of turbulence. However, high resolution atmospheric profiling instrument development is technically challenging which makes quantifying sounding data to investigate turbulence in the stratosphere very rare.

2.3 Sounding Data Analysis

The term “sounding” is used in meteorology to describe the change of temperature, and dew point with respect to altitude. In general, data from the vertical profiling of the atmosphere are used to determine atmospheric stability, precipitation type, and to determine cloud altitudes for a given time. The primary source for these data is the balloon radiosonde, an instrument package that measures temperature, relative humidity, and atmospheric pressure. These data are transmitted and the radiosonde is not recovered.

Radiosonde data frequently measure relative humidity values of 100% due to the sensitivity and response time of the humidity sensor. Nevertheless, if the difference between the temperature and dew point temperature is less than 5 degrees Celsius, clouds are assumed present (particularly in the troposphere). Data from a radiosonde observation is plotted on a chart called a "Skew-T log P" that provides a wealth of information concerning the state of the atmosphere. Radiosonde observations have a wide range of applications including:

• Input for computer-based weather prediction models,
• Local severe storm, aviation, and marine forecasts,
• Weather and climate change research,
• Input for air pollution research.

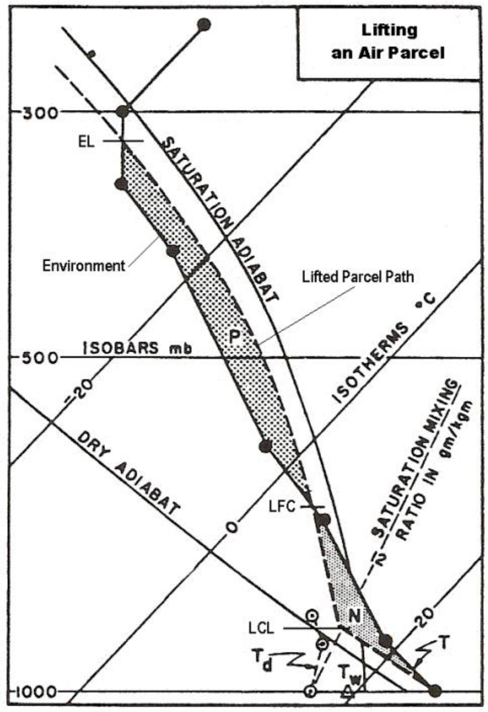

Figure 3: U.S. Air Force Example of Lifting Air Parcel on a Skew T log P chart

Figure 3 shows a typical presentation of sounding data on a plot called “skew-T, log p” diagram; the diagram features include ambient data (temperature and dew point temperature), wind data, and lapse rate which can be described in several terms listed below

• Normal lapse rate: associated with the normal decrease of temperature with altitude within the troposphere.
• Isothermal lapse rate: this lapse rate is equal to zero.
• Inversion: when the temperature increases with increasing altitude, such an inversion can be caused by fronts, subsiding air in a layer, surface radiation, or by turbulence.
• Adiabatic lapse rate: associated with the adiabatic process and it is categorized into dry and moist adiabatic lapse rates.
• Super adiabatic lapse rate: this is a positive lapse rate with a value greater than the dry adiabatic lapse rate value.

2.4 Skew T log P Diagram

These diagrams were developed so that operational forecasters and research meteorologist could move air parcels upward and downward to simulate atmospheric motion and evaluate factors such as stability without the need to do any actual thermodynamic calculations. The area on thermodynamic charts is proportional to energy. The charts typically display isobars, isotherms, mixing ratio, and a set of curves that represent changes due to dry and saturated adiabatic movement. The skew-T log p diagram was developed by the U.S. Air Force and is currently used by the U.S. military and the National Weather Service. The x-axis in the stew-T log p diagram is not orthogonal to the y-axis; rather it is rotated by 45 degrees clockwise to allow better data presentation.

In addition to the diagram mentioned above, other diagrams have been developed and have been used both operationally and in research, such as, Stuve diagram (pseudo-adiabatic chart), Aerogram (previously used by U.S. Navy, and Emagram (orthogonal skew-T log p). In this research study, the stew-T log p is adopted for sounding data analysis and conclusions. In addition to temperature and pressure lines, other lines on the diagram represent mixing ratio (ratio of water vapor density in air parcel to the density of dry air) as a measure of air humidity.

2.5 Atmospheric Stability

The stability of the vertical column is used when forecasting thunderstorms. Assessing this stability is a major part of the forecasting process. In a physical system, instability refers to a magnification or distortion of the initial state when a small disturbance or perturbation is introduced into the system. Essentially, the unstable system moves away from its initial state when perturbed. Another approach to evaluating atmospheric parcel stability is to examine the local lapse rate and compare it to the adiabatic lapse rate. The stability of a given layer is defined as:

- **Unstable**: The lifted parcel is warmer than the environment, it will Accelerate upward; this is considered an unstable condition.
- **Conditionally stable**: The lifted parcel temperature is the same as the environment; it will remain at that level; this is considered a neutral stability situation.
- **Stable**: The lifted parcel is cooler than the environment, it will sink back to its original level; this is considered stable condition.

For the unstable case, both the saturated and unsaturated parcels lifted adiabatically from the base of the layer will always be warmer than any temperature in the layer. For the stable case, both saturated and unsaturated parcels lifted adiabatically from the base of the layer will always be cooler than any temperature in the layer. For the conditional instability case, a saturated parcel lifted adiabatically from the base of the layer will be warmer than any temperature under that layer, while an unsaturated parcel lifted adiabatically from the base of the layer will be cooler than any temperature in that layer. Hence, atmospheric layer stability depends on the saturation level within the layer. Operationally the best way to evaluate instability for a location is to lift an air parcel upward and assess whether the parcel becomes warmer than the surrounding environment; which can be done by manually moving air parcel around on a sounding plot such as, Skew-T log p diagram.

2.6 Stability indices

Another method for determining atmospheric instability is the use of a stability index; which is an algorithm designed to evaluate the stability properties of a vertical sounding. Over the years numerous indices have been developed, such as, the Showalter Index, the Lifted Index, the K Index, and SWEAT. A typical automated sounding program used by the National Weather Service has more than two dozen stability index values. Some indices are:

• **CAPE:** Convective Available Potential Energy incorporates information from the entire sounding into its index value. CAPE represents the buoyant energy (positive area) between the level of free convection (LFC) and the equilibrium level (EL).

• **CIN**: Convective Inhibition energy represents the negative area below the LFC. The CIN value corresponds to the amount of energy needed to get the surface air lifted to the LFC to produce free convection.

Chapter 3: Data Analysis Methods

3.1 Convective Available Potential Energy and Convective Inhibition Energy

CAPE is defined as the integral of the positive temperature difference between an idealized rising air parcel T_{vp} and its environment T_{ve} within two specific height levels multiplied with the gas constant of dry air (Emanuel, 1994).

$$\mathrm{CAPE} = \int_{\mathrm{LFC}}^{\mathrm{LNB}} R_{\mathrm{d}} \left(\mathrm{T}_{\mathrm{vp}} - T_{\mathrm{ve}} \right) dln(p) \tag{2}$$

When moisture is present in an air parcel, the virtual temperature, Tv is used. The air parcel rises dry adiabatically from the surface to the lifting condensation level (LCL). Above the LCL, the parcel rises pseudo-adiabatically. CAPE is calculated between the level of free convection (LFC) and the level of neutral buoyancy (LNB). The equivalent potential temperature θ_{ep} remains constant during a pseudo adiabatic ascent and, therefore, is used to calculate the temperature of the rising parcel (Emanuel, 1994).

$$\Theta_{\mathrm{ep}} = T \left(\frac{p_{\mathrm{sfc}}}{p} \right)^{0.2854(1-0.28r)} exp\left[r(1+0.81r) \left(\frac{3376}{T_{\mathrm{LCL}}} - 2.54 \right) \right] \tag{3}$$

with the mixing ratio r of dry to moist air and the temperature TLCL of the parcel at the LCL, where the ascent of the air parcel changes from dry adiabatic to pseudo adiabatic. The saturation temperature T_{LCL} is approximated (Bolton, 1980) by

$$T_{\mathrm{LCL}} = \frac{2840}{3.5logT - loge - 4.805} + 55 \tag{4}$$

Large CIN does not essentially indicates strong convection; the virtual air parcel must overcome a stable layer between the surface (SFC) and LFC. The strength of this stable layer is defined by CIN. The temperature difference between the same rising air parcels as in CAPE is calculated between SFC and LFC (Williams and Renno, 1993).

$$\mathrm{CIN} = \int_{\mathrm{SFC}}^{\mathrm{LFC}} R_{\mathrm{d}} \left(T_{\mathrm{ve}} - T_{\mathrm{vp}} \right) dln(p) \tag{5}$$

As CIN defines the energy of a parcel needed to release the available CAPE energy above the LFC and therefore to be able to develop convection, CIN describes the limiting factor, which is able to prevent convection even though very high values of CAPE exist.

3.2 Theoretical description of turbulent flows

Turbulence is omnipresent, since most flows in nature and engineering applications are turbulent *[e.g. Tennekes and Llumley, 1985]*. However, after decades of turbulence investigations, a comprehensive understanding of this phenomenon is still missing and it continues to be one of the main issues in physical researches *[e.g. Frisch, 1995; Pope, 2006].* Turbulence measurements in the atmosphere are analyzed based on theoretical assumptions that are described briefly below. Since the main focus lies on the description of theories and terminologies relevant for the analysis of turbulence measurements with my developed instrument package. General review of turbulence theories can be found in standard literature [*e.g. Hinze, 1959; Tennekes and Lumley, 1985; Lesieur, 1997; Pope, 2006]*.

Due to the complexity of turbulent flows, it is not feasible to provide a precise definition of turbulence itself. Alternatively, a list of the most important characteristics of turbulence is commonly utilized. Accordingly, turbulence can be described as an unpredictable and chaotic flow-motion in space and time leading to the formation of eddies. Turbulence is initiated when internal forces in the flow field prevail the viscous forces. In addition to being dissipative, turbulence is an effective transport for momentum, kinetic energy, and heat, and it commonly extends over a wide scale range. In addition, other parameters are used in determining the transient point from laminar to turbulent flows as well as describing the turbulent flow itself. Some of these parameters are relevant for understanding turbulence theory and for measurement analysis.

As described above, turbulence appears when internal forces overcome viscous forces. The ratio between internal and shear forces is described by a dimensionless number the so called Reynolds number given below

$$Re = \frac{\text{Inertia Forces}}{\text{Viscous Forces}} \tag{6}$$

where ư, L, μ are the velocity, characteristics length, and the dynamic viscosity respectively. A laminar flow is stable as long as the Reynolds number does not exceed a critical value (critical Reynolds number R_{ecr} reaches 5×10^5 for flat plate). When Re is greater than R_{ecr}, the flow motion becomes unstable and turbulence is initiated.

Within the troposphere layer, the observed Reynolds number exceeded its critical Reynolds number value; turbulence with different intensities is found to exist in this region of the atmosphere. Reynolds number decreases with altitude due to the increase of viscous forces that becomes more dominant above a certain altitude where turbulent fluctuations can be damped efficiently.

Theoretically, turbulence can also be described using the turbulent kinetic energy equation TKE, which contains terms related to turbulence generation; such as, mechanical and buoyant forces, friction dissipation and redistribution by transport and pressure forces. Considering the relation between buoyant and mechanical production, another important turbulence parameter, the Flux Richardson number is defined:

$$Ri_{\mathrm{f}} = \frac{\overline{w'T'}\ (\frac{g}{T})}{\overline{u'w'}\ \frac{\delta \overline{u}}{\delta z}} \tag{7}$$

where u' and w' are the fluctuating horizontal and vertical wind components, T' is the temperature fluctuations, T and u are the mean temperature and the mean horizontal wind. Due to difficulties in obtaining the flux Richardson number, the gradients of potential temperature and horizontal wind are usually assumed to be proportional to the mean flux of temperature $\overline{w'T'}$, and momentum $\overline{u'w'}$ *[Tennekes and Lumley, 1985]*; therefore, the gradient Richardson number is used instead:

$$Ri = \frac{\frac{g}{\theta}\ \frac{\delta\theta}{\delta z}}{\left(\frac{\delta \overline{u}}{\delta z}\right)^2} = \frac{N_{\mathrm{B}}^2}{S^2} \tag{8}$$

Where N_B is the Brunt-Vaisala frequency and S is the wind shear.

A common assumption is that if Ri becomes negative, the production of turbulent kinetic energy increases and the atmosphere is considered statically unstable. Whereas positive values of Ri indicate that the kinetic energy is lost and the atmosphere becomes stably stratified. Turbulence will be completely suppressed, when positive Ri gets large enough *[Tennekes and Lumley, 1985]*. More precisely, from linear theory it is suggested that turbulence cannot be maintained above a critical Richardson number R_{ic} = 1/4. Only if Ri becomes smaller than R_{ic}, the mechanical production is intense enough to maintain turbulence in a stable layer *[Holton, 2004]*. Nevertheless, once created, turbulence may sustain up to Ri ~ 1. However, the existence of such a critical Richardson number has recently been questioned *[e.g. Achatz, 2005, 2007; Galperin et al., 2007; Balsley et al., 2008]*. Observations have shown that turbulent motions occur far beyond any critical Richardson number predicted by linear theory [e.g. Mauritzen and Svensson, 2007]. Atmospheric profiling data from my High Altitude Balloon experiments will be used and the existence of a critical Richardson number R_{ic} will be investigated.

3.3 Atmospheric Profiling System

The governing equations also called, the primitive equations, consist of the equation of state (ideal gas law), the conservation of mass (continuity equation), the conservation of momentum (Newton's second law), the conservation of moisture, the conservation of heat (first law of thermodynamics), and the conservation of scalar quantity [Holton, 1992]. Measurements of the vertical profiles of temperature, humidity, and wind are indispensable to monitor and understand physical processes in the atmosphere and to verify, evaluate, and improve weather prediction models. By measuring these variables, the stability of an atmospheric layer or a region can be determined and mean fluxes of heat and momentum can be estimated.

My developed integrated meteorological profiling system used in this research, is a sensor package designed to measure and record atmospheric data to altitudes reaching 30 km. The instrumentation package consists of pressure, temperature, humidity, and light sensors as well as a wind meter and Global Positioning System unit (GPS). The integrated system allows the user to communicate commands to the sensors through its microcontroller unit, which also gives the user the option to modify, upgrade, or change the system inputs and outputs for specific data analysis. The system uses a software program called Arduino written in C^{++} language, which allows the user to control the sensors operation criteria such as changing the frequency of the measured data as well as selecting the output or calculated values to be used by atmospheric simulation models and analysis.

3.4 Sounding Instrumentation

The limitation of manned airborne atmospheric measurements at high altitude has led to the development of an integrated profiling system for the purpose of measuring and recording atmospheric sounding data within the troposphere and lower stratosphere. The instrument was tested and verified for accuracy and reliability. The restrictions posed by the Federal Aviation Administration on the payload weight was the main constraint in choosing the right components for in this project. The system was designed to be reusable, have long-period maintenance occurrence, and short recalibration time, which can lead to lowering the cost of operating and maintaining the system. The instrumentation package was designed to facilitate easy recalibration or replacement of its components upon failure.

The atmospheric profiling system is a necessary component of the project because it allows achieving one of the main objectives of this research which is to measure the inherent vertical stability of the stratosphere layer (above the conventional aircraft ceilings) and to investigate the turbulence characteristics observed in the stratosphere. These airborne measurements of the meteorological parameters in the middle stratosphere were used to investigate the stability and turbulence in the atmosphere as well as to parameterize and validate the use of existing atmospheric numerical models and methods in predicting turbulence.

3.5 System Integration and Performance

The integrated meteorological profiling system demonstrated to be capable of performing atmospheric measurements of temperature, humidity, pressure, and wind velocity using commercially available solid-state sensors. The sensors have high accuracy, resolution, and the ability to perform in harsh environment conditions, such as, low temperature, low pressure and in the stratified regions where the relative humidity can reach 100%. In addition, a data logger unit for recording atmospheric measurements was integrated into the system. The system allows multiple measured data to be serially streamed into a PC; the identification of data from different sensors is accomplished by including a microcontroller unit (open-log).

The meteorological system was used as the primary payload for the balloons. The components of this payload consist mainly of pressure, temperature, wind and humidity sensors, however, the system has the ability to integrate other sensors such as ultraviolet light and solar radiation. In addition, this meteorological system allows the user to communicate commands to the sensors using its microcontroller unit, which also gives the user the option to modify, upgrade, or change the systems inputs and outputs for specific data analysis. The system uses a computer program (an open source code, written in C language) which allows the user to regulator the sensors operation criteria, such as, controlling the frequency of data measurements so that the outputs can be readily integrated into existing numerical atmospheric models.

3.6 Sounding System Components

The Weather Shield is an Arduino shield that measures barometric pressure, relative humidity, luminosity and temperature as well as wind speed, direction, and GPS unit.

This micro-Shield equips the Arduino with mass-storage capability to be used for data-logging or other related projects. Communication with micro-SD cards is achieved over an SPI interface.

The Red-Board is programmed over a USB Mini-B cable using the Arduino IDE, and it uses an open source code (written in a C language) which was modified to include the calculate and analyze different input and output data for our different parameters. Red-Board has 14 Digital I/O pins, and 6 Analog Inputs.

Chapter 4: Flight Test Method

4.1 Flight Test Method and Results

We used High-altitude balloons (HABs) to collect weather data, including pressure, temperature, humidity, and wind speeds, up to an altitude of ~ 100,000 feet (30km). The high altitude balloon experiments conducted by me at the University of Tennessee Space institute in Tullahoma had the objective of developing and flying low cost atmospheric profiling system and the analysis of the resultant data. An overview of two flights conducted in January and May of 2017 are below. A description of the launch procedures, payload, method of tracking and recovery, and predicted trajectory of the balloon are presented here. This project to study turbulence in the stratospheric layer was made possible through the development of low cost scientific payloads and flight test methods, and, procedures for carrying out these scientific experiments at minimal costs.

The purpose of my project was to conduct a series of high altitude balloon experiments to obtain measurements and wind fluctuations within the troposphere and lower stratosphere. Sounding data obtained from the launches with Westwood Middle and R. E. Lee Elementary are used in the data analysis for this research. The measurements were: temperature, relative humidity, pressure, GPS location and relative horizontal wind. In addition, a GPS tracking unit was also included as one of the system's component to facilitate the retrieval of the payload.

The latitude, longitude, and altitude data recorded by the GPS was used to plot the balloon's trajectory in a three dimensional coordinate system. The Westwood HAB was launched on 01/20/2017 of from the city of Manchester in Tennessee and landed near the city of Towson, TN. The flight test duration was about 2 hours and 75 minutes and covered a ground distance of 210 km. The figure below is a three-dimensional flight trajectory for this experiment

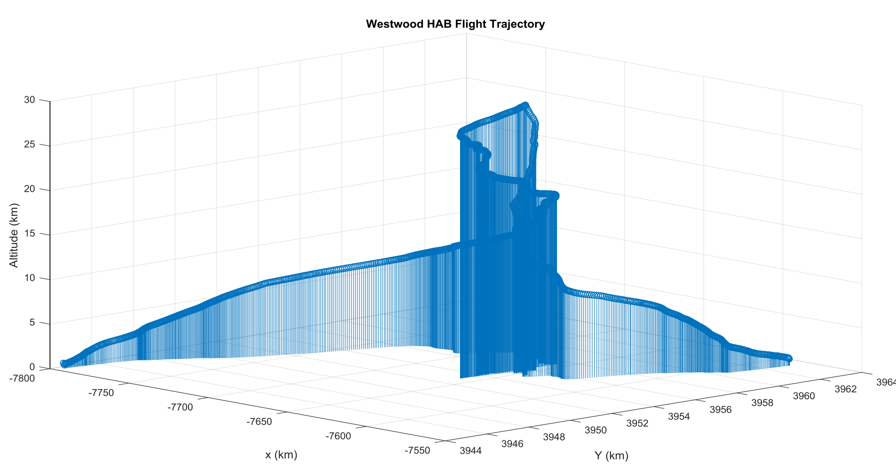

Figure 4: High Altitude Balloon Trajectory for Westwood Flight Test

The latitude, longitude, and altitude data recorded by the GPS was used to plot the balloon's trajectory in a three dimensional coordinate system. The R. E. Lee HAB was launched on 05/20/2017 from the city of Tullahoma in Tennessee and landed near the city of Crossville, TN. The flight test duration was about 3 hours and covered a ground distance of 140 km. The figure below is a three-dimensional flight trajectory for this experiment. Partial data from the GPS unit was missing during the descent phase (showing as a gap in figure)

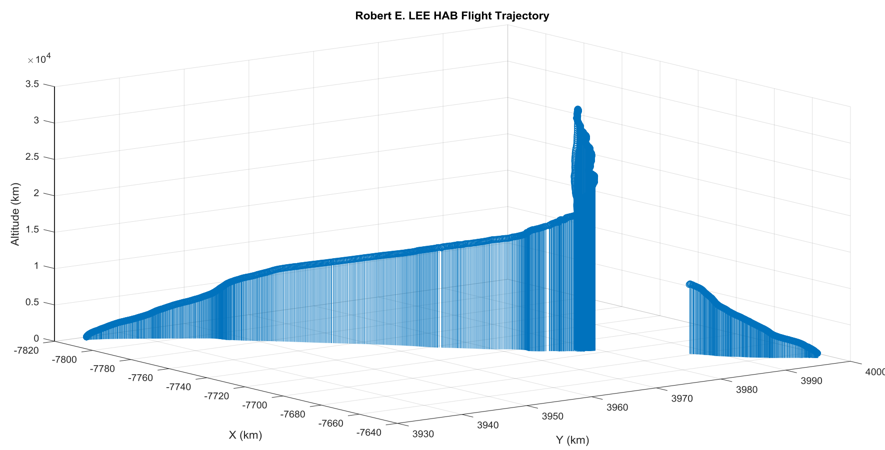

Figure 5: High Altitude Balloon Trajectory for Robert E. Lee Flight Test

4.2 Atmospheric Profile

Measured environmental temperature and pressure profiles, with respect to altitude, are plotted in the figure below, however; air density profile was approximated using equation 1: Where ρ is air density in $\left(\frac{kg}{m^3}\right)$, T is temperature in kelvin, and R is the gas constant for dry air and is equal to $287\ \frac{J}{(kg)(K)}$.

The figure shows a decrease in the atmospheric temperature, pressure, and density with increasing altitude, up to about 15 km, which marks the edge of the troposphere region. After the 15 km altitude, the atmosphere temperature, pressure, starts to increase with altitude, which marks the beginning of the stratosphere layer. Recorded atmospheric data was limited by the HAB burst altitude of about 30 km.

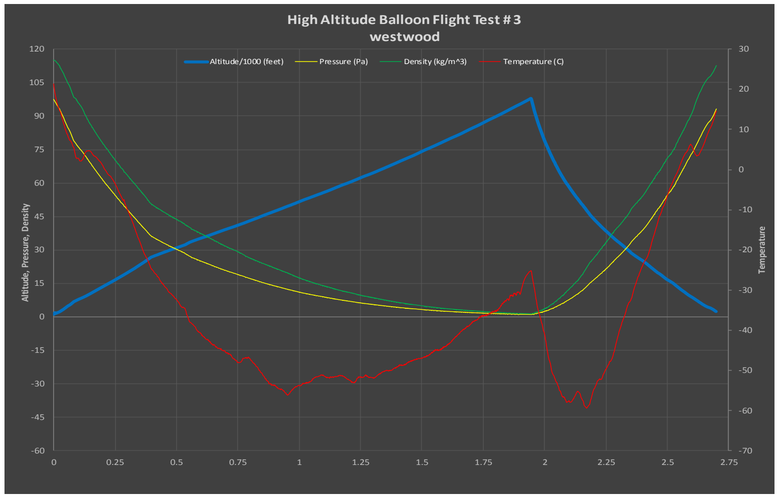

Figure 6 : Atmospheric Profile of Temperature, Pressure, density, and Altitude versus time

As mentioned above the standard atmospheric conditions are used only for comparison reference . The standard atmosphere layers are devided based on the temperature structure (lapse rate) for standard and it is ploted versus the measured envirnemental temperature in the figure below.

The major difference between the standard and the measured temperatures for this flight test is the absence of the tropopause layer (an isothermal layer). While the standard temperature lapse rate within the troposphere is about the same as that of the envirementaI temperature it shows some consistency within the lower region of the stratosphere region.

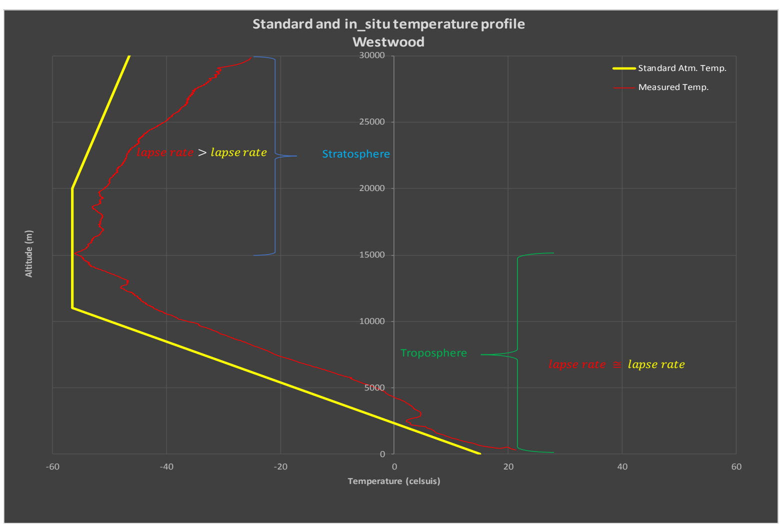

Figure 7: In-Situ and Standard Atmospheric Temperature Profiles versus Altitude

The figure below shows a decrease in atmospheric temperature, pressure, and density with increasing altitude, up to about 13 km, which marks the edge of the troposphere layer and the beginning of the tropopause layer where the temperature is relatively shows to be constant. At 18 km altitude, the atmosphere temperature and pressure starts to increase with altitude, which marks the beginning of the stratosphere layer. Recorded atmospheric data was limited by the HAB burst altitude of about 30 km

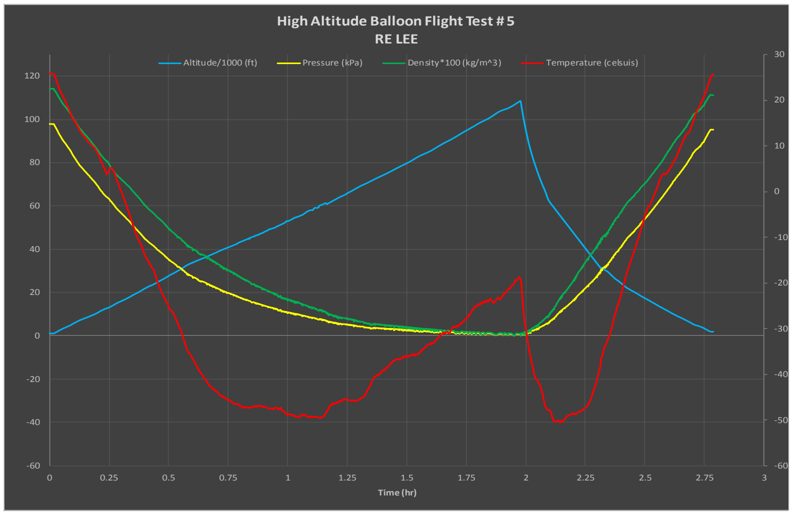

Figure 8: Atmospheric Profile of Temperature, Pressure, density, and Altitude versus time

The figure below shows that within the troposphere, the temperature lapse rate for standard atmospheric conditions is consistent with the measured temperature lapse rate. The figure also shows the presence of an isothermal layer (tropopause) between 13 km and 18 km. However, within the lower region of the stratosphere. the temperature lapse rate was found to be slightly greater than the standard lapse rate

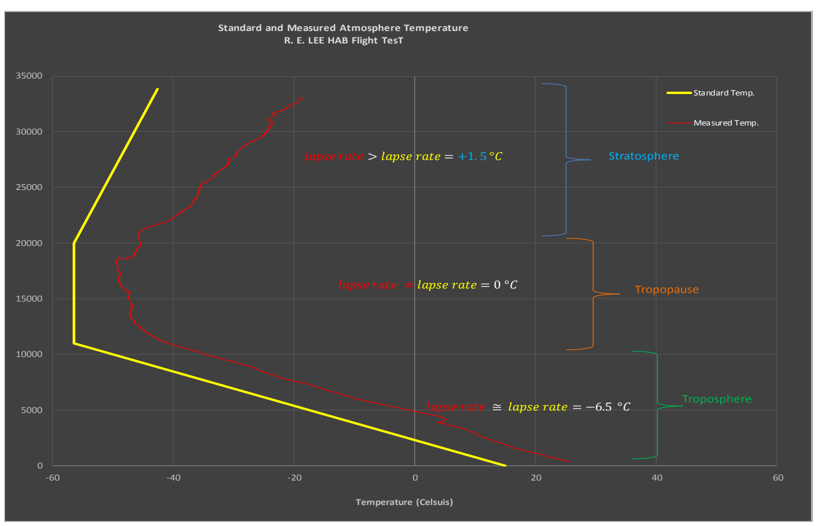

Figure 9: In-Situ and Standard Atmospheric Temperature Profiles versus Altitude

4.3 Mean and Turbulent Wind Gust

The following graphs show the mean horizontal wind (blue) and the measured relative wind (turbulence perturbation or gust velocity; green) from the R. E. Lee ascent data. The balloon during ascent rises at a uniform rate of 4.1 m/s. In addition, the balloon is advected horizontally at the speed of the mean wind (blue line). In a Lagrangian frame of reference the horizontal wind speed sensor (green line) represents the turbulent perturbation, or gust, superimposed on the mean wind. My interpretation of each graph is given below the graph. As shown above the jet stream is located from 9,000 to 17,000m. Turbulence is non-zero below 25,000m and peaks at 18,000m.

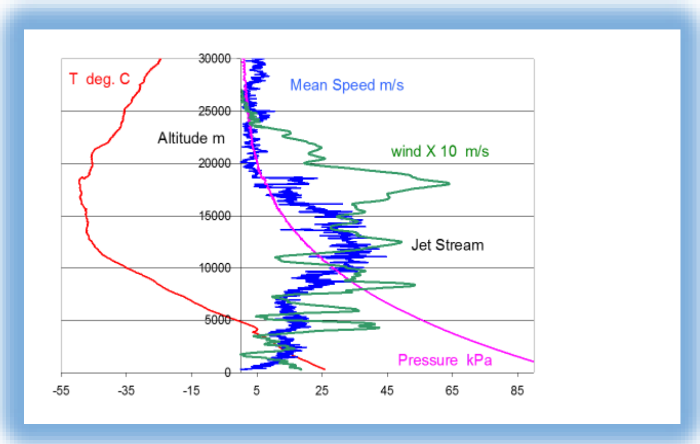

Figure 10: Mean and wind speeds versus Altitude from Westwood HAB Experiment

The Peak turbulence at 18,000m corresponds to a step change in the mean velocity near the top of the jet stream layer. This also corresponds to an abrupt change in temperature within a length of ~1km. This indicate a very large single eddy was encountered near the top of the jet stream.

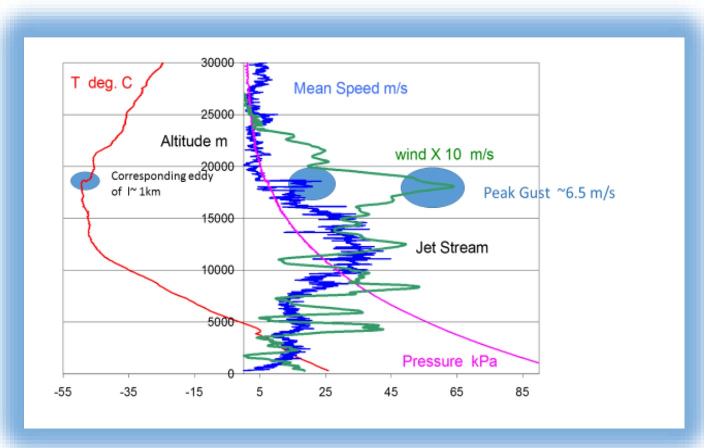

Figure 11: Wind Gust and wind speed from Westwood HAB Experiment

From the video included in the payload, the lower peaks of turbulence at 4,000m and 6,000m correspond to cloud layers. Within clouds condensation releases the heat of vaporization to the surrounding air parcel causing a temperature increase and a jump in buoyancy which increases the turbulence level. This can be felt on airliners as turbulence increases as they ascend or descend through cloud layers.

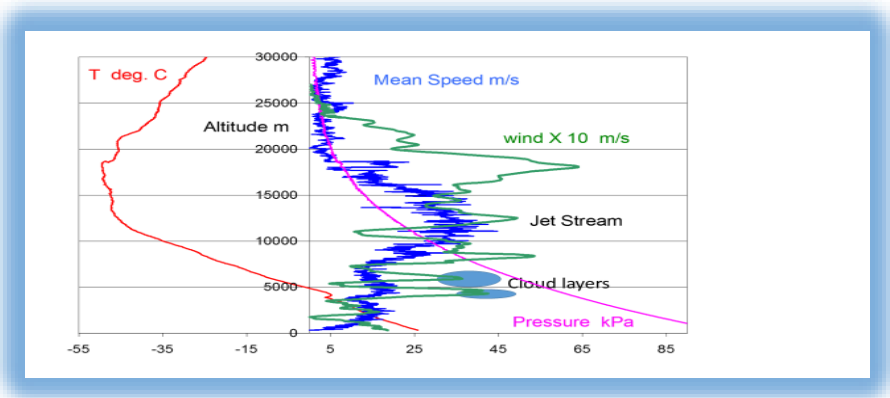

Figure 12: Mean and wind speeds versus Altitude from R. E. Lee HAB Experiment

The highest turbulence peaks occur at the mean velocity inflection points entering and exiting the jet stream. The Rayleigh theorem in fluid mechanics states that an inflection point in a velocity profile is a necessary a prerequisite for turbulence. While this theorem has been shown to have exceptions, peak wake turbulent does occur at the inflection points of the mean flow. My results are consistent with peak turbulence at the inflection points of the mean flow. This region of the lower jet stream is well known to airline pilots as a location of clear air turbulence. Here the peak turbulence at 18,000m would never be encountered by a commercial airliner because their highest ceiling is ~15,000m.

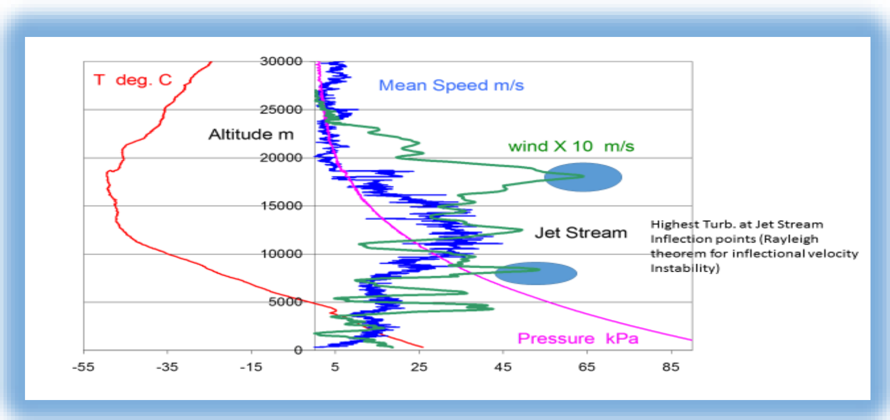

Figure 13: Wind Gust and wind speed from R. E. Lee HAB Experiment

Chapter 5: Results and Conclusions

5.1 Richardson number

My visual interpretation of the data was given in the previous section. The Richardson numbers and stability conditions give an objective determination and classification of the stability regime as a direct approach to the numerical examination of atmospheric turbulence. The establishment of stability criteria is commonly accomplished by use of the non-dimensional Richardson number; which is quite accurate if certain precautions are observed in its calculation. It is necessary to understand the turbulent process within the atmospheric layer in question in order to be able to use the Richardson number effectively as an identifier of the stability regime. Because the numerical calculation of the Richardson number is highly dependent on the vertical gradient of wind velocity and temperature, accurate measurement of these parameters are essential in terms of whether the data are representative or have been biased.

The Richardson number is a nondemisional parameter used in the determination of the characteristics of dynamic similarity according to Batchelor (1953), it is the most useful stability indicator in most study related to atmospheric turbulence analysis. Richardson derived a ratio of work done against gravitational stability to energy transformed from laminar to turbulent motion while investigating the effects of gravity on the suppression of turbulence.

The gradient Richardson number, R_i, has largely been used as a criterion for assessing the stability of stratified shear flow. It is defined as

$$Ri = \frac{-g\frac{\partial \rho}{\partial z}}{\rho\left(\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right)}, \qquad (9)$$

The negative sign is to guarantee positive values for stable stratification condition. Large values of R_i indicate a very stable condition, while low values may indicate the presence of dynamic instability.

By definition, N is the buoyancy frequency (Brunt Vaisaila frequency) is

$$N = \left(-\frac{g}{\rho}\frac{\partial \rho}{\partial z}\right)^{1/2} \qquad (10)$$

And S is the total vertical wind shear

$$S = \left[\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right]^{1/2} \qquad (11)$$

Therefore the gradient Richardson number can be expressed as

$$Ri = \frac{N^2}{S^2} \quad (12)$$

This equation indicates values R_i is directly proportional to $N^2$and indirectly proportional to S^2. The common interpretation is that low stratification (small N) leads to subcritical R_i values which are indicative of unstable conditions, while large values of the buoyancy frequency indicate high stratification because of the strong restoring buoyancy force which prevents mixing. While, small values of the Vasaila frequency are indicatives of low static stability, dynamic stability which is determined by R_i depends on the vertical wind shear as well.

5.2 Stability Classification Scheme

In 1961, Frank Pasquill offered a scheme for categorizing atmospheric stability based on available surface observations such as wind speed, cloud cover, and solar radiation. Stability is classified into six categories: extremely unstable (A); moderately unstable (B); slightly unstable (C); neutral (D); slightly stable (E); and moderately stable (F). In this thesis, the performance of the Pasquill scheme is tested in order to determine whether it is appropriate to use for detecting the presence and intensity of atmospheric turbulence. These stability categories range from A (extremely unstable) to F (moderately stable), with D being the neutral category listed in table 1.
The Businger version of Richardson number was developed by Sedefian and Bennett (1980) and defined in finite difference form by Equation below

$$Ri = \frac{g}{T(z_1)} \frac{\left[\frac{T(z_1)-T(z_2)}{z_1-z_2}\right]}{\left[\frac{u(z_1)-u(z_2)}{z_1-z_2}\right]^2} \quad (13)$$

Table 1: Pasquill Stability Class Correlation with Richardson Number

Atmosphere Stability Class Correlated with Richardson Number

Pasquill Stability Class	$Ri = \frac{g(\Delta T/\Delta z)}{T(\Delta u/\Delta z)^2}$
A	$Ri < -0.86$
B	$-0.86 \le Ri < -0.37$
C	$-0.37 \le Ri < -0.10$
D	$-0.10 \le Ri < 0.053$
E	$0.053 \le Ri < 0.134$
F	$0.134 \le Ri$

The Pasquill stability scheme is only one of a number of methods used to determine pollutant concentrations, and several studies have compared the effectiveness of the Pasquill scheme against these other methods.

Robert E. Lee HAB flight test data of temperature and wind velocity (u) were used in the calculation of the Richardson number values for altitudes reaching 30 km. The pasquill clasifications criteria was inmplemented to estimate tubulence intensity within the troposphere and lower stratosphere layers; the results are ploted in the figure below. Based on Pasquill scheme, the most stable region of the atmosphere exists within the stratopshere layer, where R_i > 0.134 and labeled as category F. While within the troposphere the schame indicates the existence of tublence with intesity ranging for unstable for the most part to strongly unstable within few layers of the troposphere.

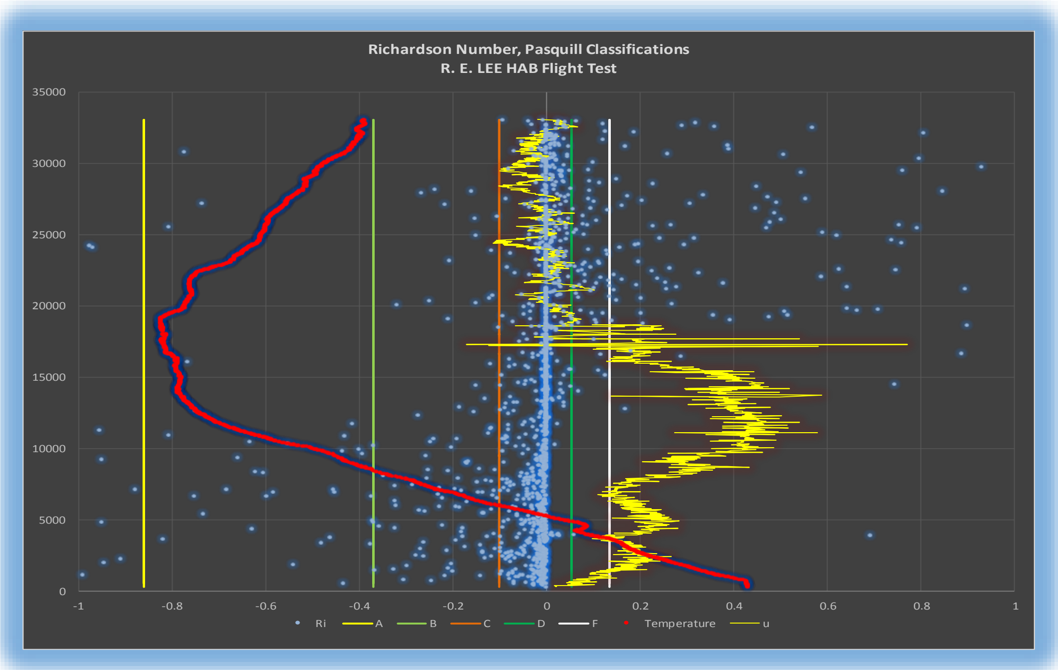

Figure 14: Pasquill Classification of Richardson Number from Robert E. Lee HAB Flight Test

Westwood HAB flight test data of temperature and wind velocity (u)were used in the calculation of the Richardson number values for altitudes reaching 29 km. The pasquill clasifications criteria was inmplemented to estimate tubulence intensity within the troposphere and lower stratosphere layers; the results are ploted in the figure below.
Based on Pasquill scheme, the most stable region of the atmosphere exists within the stratopshere layer, where R_i > 0.134 and labeled as category F, although there are several regions where the scheme indicates the existence of categories B and C for unstable wind. While within the troposphere the schame indicates the existence of tublence with intesity ranging for unstable for the most part to strongly unstable within few layers of the troposphere.

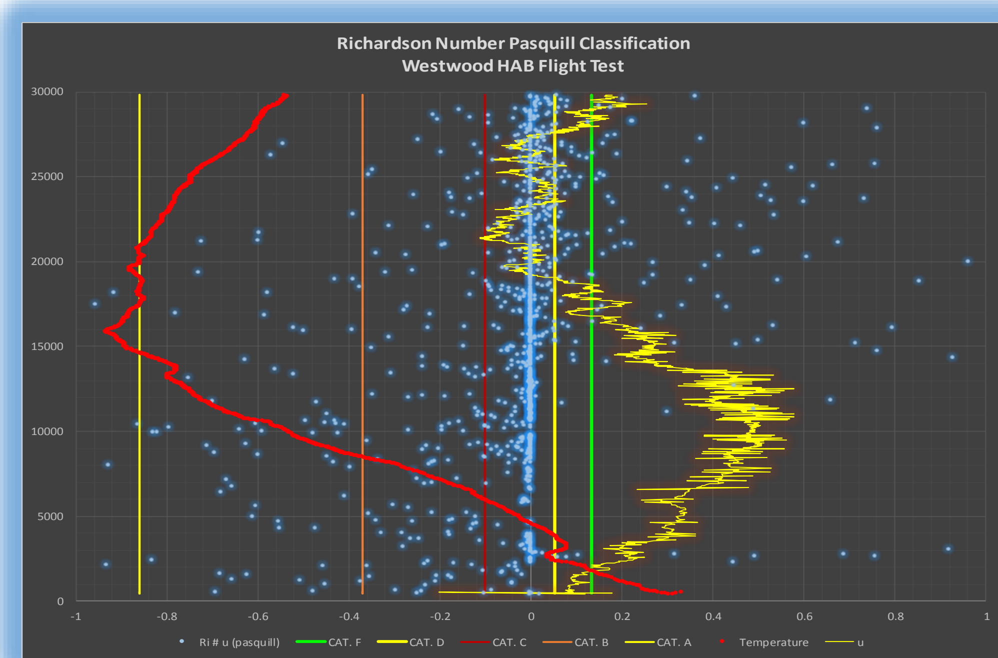

Figure 15: Pasquill Classification of Richardson Number from Westwood HAB Flight Test

Another commonly used stability classifications scheme is based on Mauristen and Svensson, shown in table 2. This method uses equation 8 to calculate the gradient Richardson number; however, the classification intervals have different values from those of the Pasquill classification and are listed in the table below:

Stability Classification	**Richardson Number**
Strongly unstable	$R_i < -0.2$
Unstable	$-0.2 < R_i < -0.1$
Neutral	$-0.1 < R_i < 0.1$
Stable	$0.1 < R_i < 0.25$
Strongly stable	$R_i > 0.25$

Table 2: Mauritsen and Svensson Stability Class Correlation with Richardson Number

Robert E. Lee HAB flight test data of temperature and wind velocity (u) were used in the calculation of the Richardson number values for altitudes reaching 30 km. Mauritsen and Svensson clasifications criteria was inmplemented to estimate tubulence intensity within the troposphere and lower stratosphere layers; the results are ploted in the figure below.

Based on this scheme, the most stongly stable region of the atmosphere exists within the stratopshere layer, where $R_i > 0.25$. While within the troposphere the schame indicates the existence of tublence with intesity ranging for unstable for the most part to strongly unstable within few layers of the troposphere where R_i is less or equall to -0.2 . However, this scheme classifications indicates more unstable regions within the troposphere as well as the lower stratosphere while the Pasquill scheme caracterized these layers to be either stable or neutral.

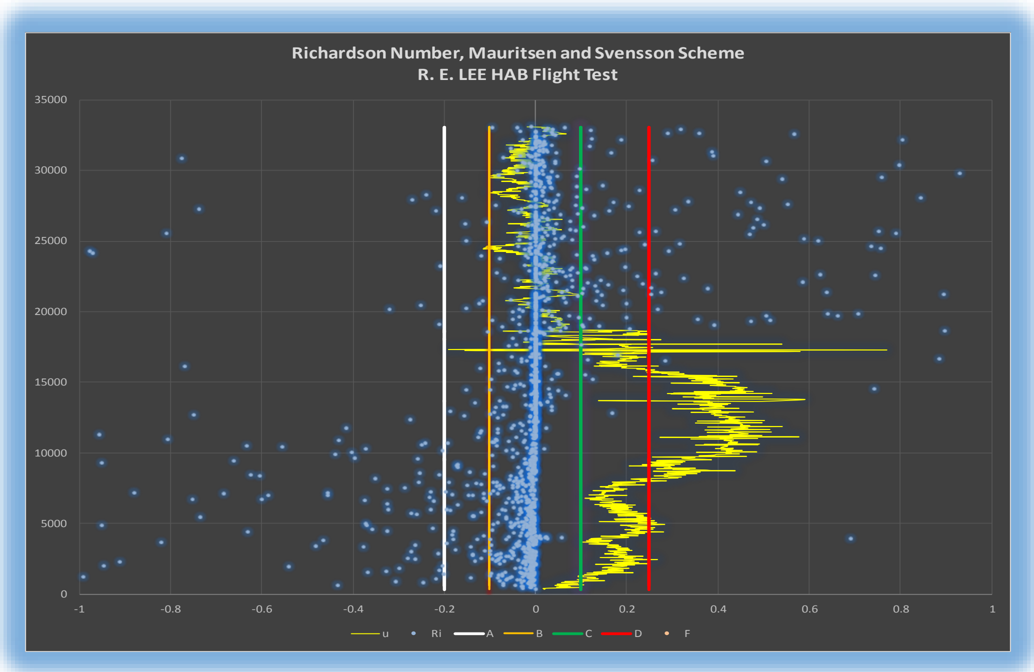

Figure 16: Mauritsen and Svensson Classifications of Richardson Number from Robert E. Lee HAB Flight Test

5.3 Bulk Richardson Number

This form Richardson number is frequently used by meteorologists to approximate wind shear and temperature gradients at discrete heights. The values of the critical Richardson number do not apply to these finite differences across thick layers. The thinner the layer, the closer the value to the theory

$$R_B = \frac{\frac{g}{\overline{\theta_v}}\frac{\Delta\overline{\theta_v}}{\Delta z}}{\left(\frac{\Delta\overline{U}}{\Delta z}\right)^2 + \left(\frac{\Delta\overline{V}}{\Delta z}\right)^2} = \frac{g\Delta\overline{\theta}_v\Delta z}{\overline{\theta_v}((\Delta\overline{U})^2 + (\Delta\overline{V})^2} \tag{14}$$

Using atmospheric, wind, and coordinates data collected during HAB flight tests, the Richardson number was calculated using the equation below

$$Ri = \frac{\frac{g}{\theta}\frac{\partial \theta}{\partial z}}{\left[\left(\frac{\partial U}{\partial z}\right)^2 + \left(\frac{\partial V}{\partial z}\right)^2\right]} \quad (15)$$

Where g is the acceleration due to gravity g = 9.81 m/s, ϴ the potential temperature, U the east-west component of wind velocity, V the north south component of the wind, and z is the height. When the negative sign (negative temperature lapse rate) is omitted from the ratio in equation; therefore, when atmospheric condition is favorable for the presence of turbulence, R_i is less than the critical Richardson number R_{ic} = 0.25. The vertical gradient of wind components are calculated using data recorded by the GPS. The potential temperature profile and its gradient was computed using ambient temperature and pressure data.

The Bulk Richardson was calculated using the potential temperature and wind velocities (u,v) recorded by the profiling system. Based on this scheme the stratosphere layer is where the indication of some stability seems to exits; that is where R_B is greater than 0.25, while within the troposphere is considered to be unstable. Also, it indicates that turbulence exists in the most parts of the lower stratosphere layer. Therefore one can conclude that for this flight test atmospheric layers within altitudes of 30 km can be considered either very unstable or unstable.

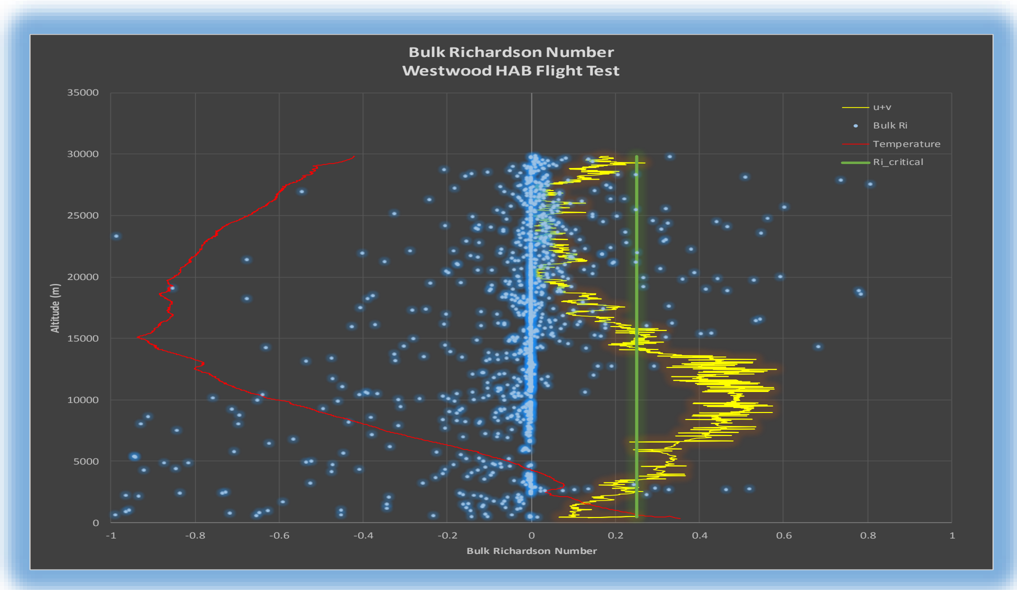

Figure 17: Computed Bulk Richardson Number versus Altitude for Westwood HAB Flight Test

For R. E. LEE flight test the stratosphere layer can be described as being fairly stable, where the R_B > 0.15 for the most part; while indication of regions with strong stability criteria is shown to exists within few layers of stratosphere where R_B is greater than 0.25. While within the troposphere region atmospheric condition for this flight test ranges from neutral to unstable when the Bulk Richardson criteria is used.

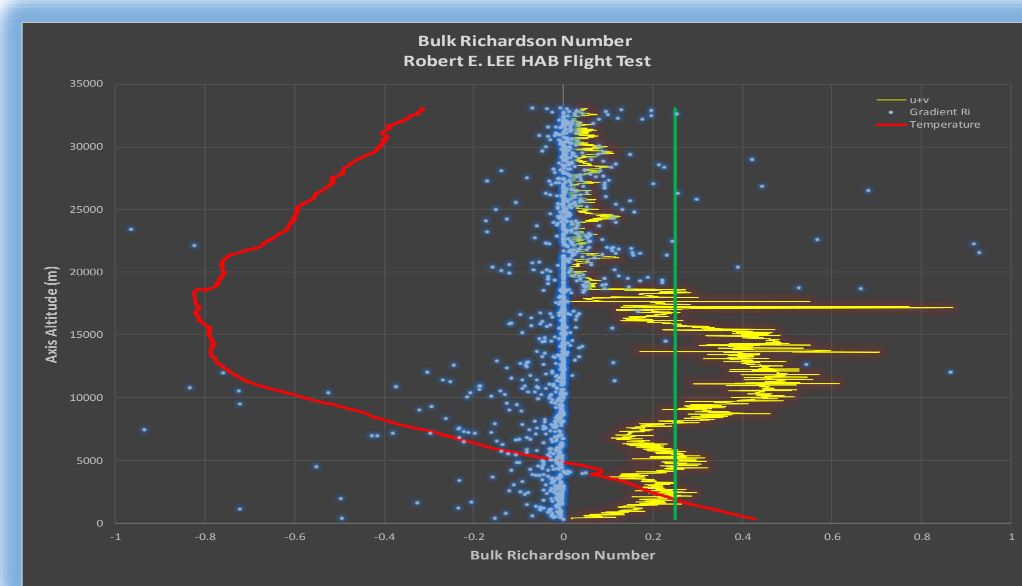

Figure 18: Computed Bulk Richardson Number versus Altitude for Robert E. Lee HAB Flight Test

5.4 Air Parcel Theory

Atmospheric soundings on Skew-T log P diagram

- Allow us to identify stability of layer
- Allow us to identify various air masses
- Tell us about moisture in a layer
- Allow us to speculate on occurring processes

Why use adiabatic diagram

- They are designed so that the area on the diagrams is proportional to energy
- The fundamental lines are straight forward and easy to implement
- On a skew T log P diagram the isotherms are orthogonal to the isentropes

The information on adiabatic diagrams allow us to determine

- Convective available potential energy, on the thermodynamic diagram CAPE represent an area that is proportional to energy. It is the area between the air parcel temperature and the environment temperature from the LFC to the EL. CAPE is a measure of instability.

- Convective Inhibition, on the thermodynamic diagram CIN also represents an area proportional to energy. It is the integrated area from the surface to the LFC level.

Critical levels on a thermodynamic diagram

- Lifting condensation level (LCL): The level at which a parcel lifted dry adiabatically will become saturated.
- Level of free convection (LFC): The level above which a parcel will be able to freely convect with no additional forcing.
- Equilibrium level (EL): The level above the LFC at which an air parcel will no longer be buoyant; at this point, the environment temperature and parcel temperature are equal. Above this level, the parcel is said to have a negative buoyancy. The parcel may continue to rise above the EL. This is called an "overshoot" due to its momentum generated by accumulated kinetic energy during its rise.
- Convective condensation level (CCL): Level at which the base of the convective clouds will begin.
- Mixing condensation level (MCL): Level at which clouds will form when a layer is sufficiently mixed mechanically due to turbulence.

The plots below were generated using a developed MatLab program. The program's methodology is based on the laws of thermodynamics. The computer code uses the sounding data as input and it calculates several parameters that are used by meteorologists to predict weather conditions. However, in this thesis the program's output will be limited to the calculation of the Convictive Inhibition Energy (CIN) and The Convective Available Potential Energy (CAPE).
The classification categories for stability using values of calculated CAPE and CIN are listed in the tables 3 and 4

Table 3: Stability Category Classifications for Convective Available Potential Energy

CAPE value (J/kg)	Stability Criteria
0	Stable
0 to 1000	Marginally Unstable
1000 to 2500	Moderately Unstable
2500 to 3600	Very Unstable
3600 +	Extremely Unstable

Table 4: Stability Category Classifications for Convective Inhibition Energy

CIN value (J/kg)	Stability Criteria
-25 to 0	weak resistance that can be easily broken by surface heating
-50 to -200	moderate resistance that can be broken by strong heating
-200 +	strong resistance that impedes thunderstorm development

The calculated values of CIN and CAPE computed using the Matlab program are listed below the Skew-T lop P plots for Westwood and Robert E. Lee flight tests respectively. Based on these values and using the stability criteria listed above, one can conclude that for HAB conducted 01/20/17 the atmospheric condition are considered marginally unstable, while for HAB test on 05/18/17, the atmosphere is said to be moderately unstable.

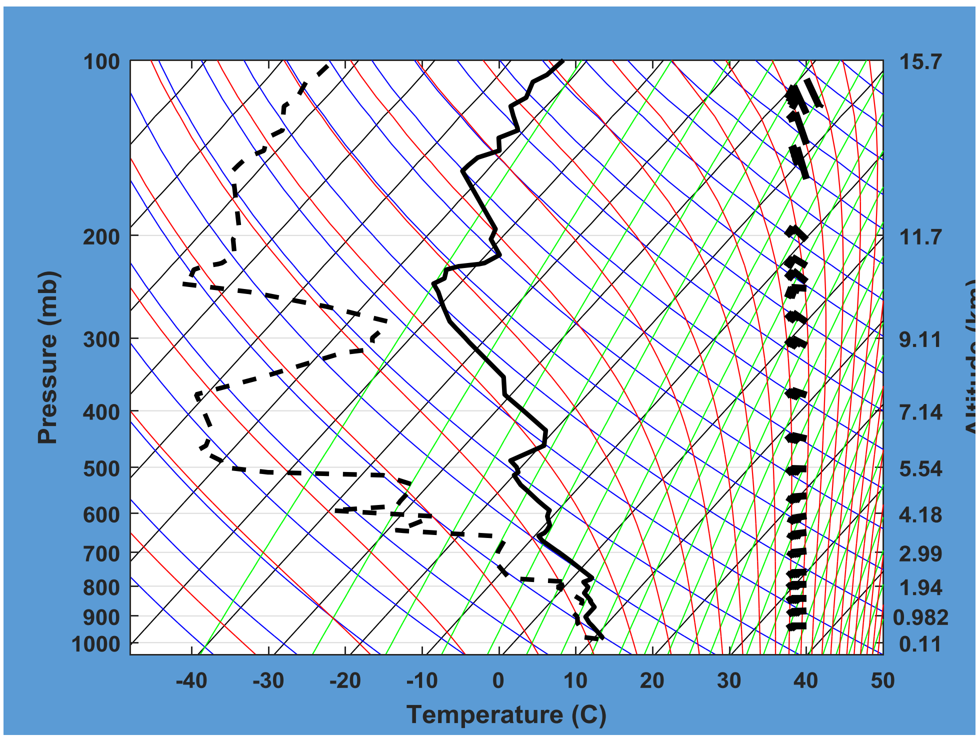

Figure 19: Skew-T Log P Diagram from Westwood HAB Flight Test

CAPE: **2.49** J/kg
CIN: **-151.18** J/kg

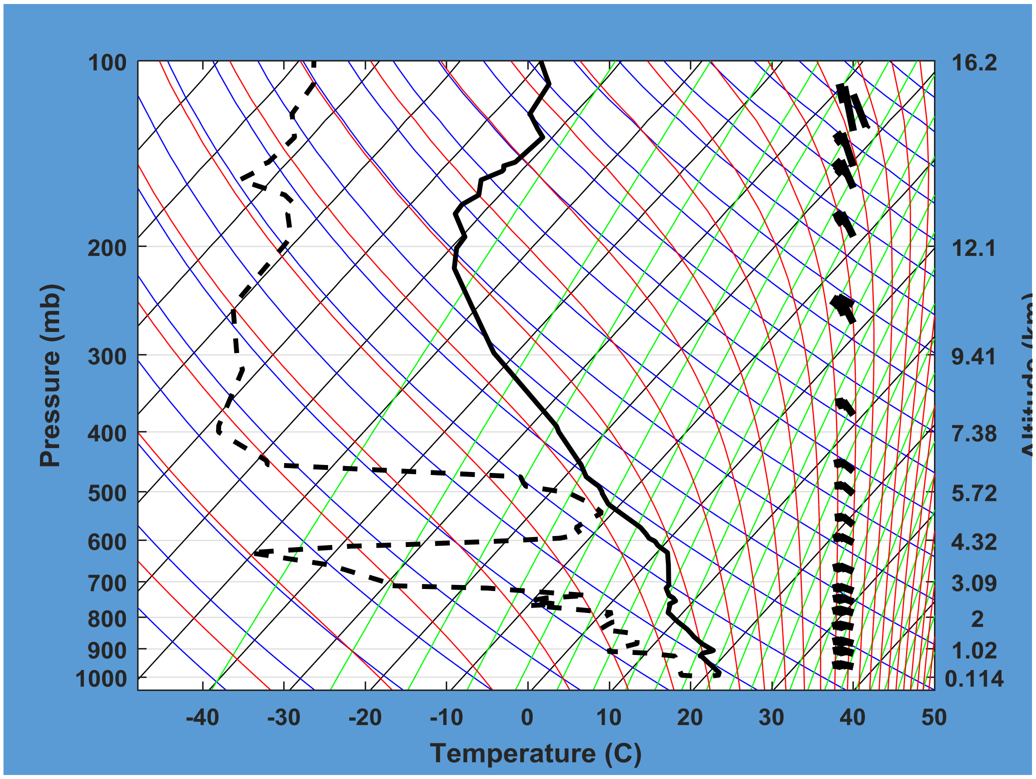

Figure 20: Skew-T Log P Diagram from Robert E. Lee HAB Flight Test

CAPE = **935.72** J/kg
CIN = **-35.68** J/kg

5.5 Results and Conclusions

The scope of this work was to develop a small-scale payload for turbulence observations in the troposphere and middle stratosphere. Using atmospheric stability methods determined via the Pasquill and Mourissen classification scheme has been tested against stability classes determined using the Bulk Richardson number and the air-parcel methods, the most used methods for stability observation in meteorology.

Results of Bulk R_i number indicated atmospheric moderate instability from HAB experiments, while Pasquill determined the stability to be neutral. Stable environments are typically taking place when the neutral category is selected; however, the Pasquill stability scheme did correctly predict neutral stability for most of the cases where neutral conditions were actually present.

In general, both of the parcel method and the Richardson number scheme predicted marginal and moderate instability within the troposphere and the lower stratosphere respectively. It is concluded that primarily, there is some degree of correlation between these methods in detecting atmospheric instability; however, they do differ in the determination of turbulence intensity classification. Also, the results from this research shows that the stratosphere layer of the atmosphere is not totally stable as considered to be due to the fact that air is stratified within this layer, in fact the results from the Richardson number indicate several layers with thickness of 30 meter or less to be very unstable.

Turbulence in the middle stratosphere has been an oft ignored topic. National Weather Service radiosonde are designed to burst at 17,000m. The FAA and airliners are unconcerned about turbulence above their 15,000m flight ceilings. Aviation weather (aviationweather.gov) includes turbulence forecasts to just 15,000m. Weather forecasters are unconcerned because even the tallest hurricane or thunderstorm only extends upwards to ~16,000m. However the future of high-speed flight includes much higher altitudes (to 30 km). Accurate assessment of expected turbulence is an important design criteria. The term stratosphere is from "strato" meaning layered or stable. Here I have shown that the stratosphere while typically more stable than the troposphere, does experience significant turbulence.

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Vita

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