1   Catalina Eddy as Revealed by the Historical Downscaling of Reanalysis

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19   **Abstract**

20   Climatological properties, dynamical and thermodynamical characteristics, and

21   long-term variabilities of the Catalina Eddy are examined from the 61 year

22   NCEP/NCAR Reanalysis downscaled to hourly 10 km resolution.   The eddy is

23   identified as a mesoscale cyclonic circulation confined to the Southern California Bight

24   region. Pattern correlation of wind direction against the canonical Catalina Eddy is used

25   to extract cases from the downscaled analysis. Validation against published cases, the

26   NCEP operational North American Mesoscale analysis, geostationary satellite images,

27   surface observations and buoy observations confirmed that the downscaled analysis

28   accurately reproduces Catalina Eddy events.

29   Large-scale features associated with Eddy events are characterized by zonal

30   wind at 500 hPa and a cutoff low positioned over southeastern California/southern

31   Nevada at 850 hPa, which enhances the northeasterly flow over Central California

32   below this level. Near the surface, the Catalina Eddy is characterized by a streak of

33   positive vorticity emanating from the coast of the Santa Barbara Channel, with two

34   centers, the southern one corresponding to the Eddy center. The eddy is associated with

35   convergence at the center and divergence along the coast. The temperature anomaly is

36   warm to the west and cold to the southeast. A positive humidity anomaly is apparent to

37   the east of the eddy, extending well into land. A vertical cross section indicates a

38   shallow vorticity structure reaching only up to 900 hPa. The thermal structure indicates

39   an eastward tilt with height.

40   A composite analysis of the initiation phase of the eddy indicates that no

41   apparent large-scale cyclonic/anti-cyclonic large-scale forcing is associated with the

42   eddy formation or decay. The source of the vorticity is located at the coast of the Santa

43   Barbara Channel. It is generated by the convergence of the wind system crossing over

44   the San Rafael Mountains and the large-scale northwesterly flow associated with the

45   subtropical high. This vorticity is advected towards the southeast by the northwesterly

46   flow which contributes to the formation of the streak of positive vorticity. At 6 hours

47   prior to the mature stage, there is an explosive generation of positive vorticity along the

48   coast, coincident with the phase change of the sea breeze circulation (wind turning from

49   onshore to offshore), resulting in the convergence all along the California coast. The

50   generation of vorticity due to convergence along the coast together with the advected

51   vorticity from the north resulted in the formation of southerly flow along the coast,

52   forming the Catalina Eddy. An examination of the surface pressure anomaly evolution

53   showed that vorticity is leading the eddy formation rather than the pressure gradient,

54   since no apparent formation of a south-to-north pressure gradient along the coast at the

55   initiation phase is recognized. Seven to eight hours prior to the mature stage, a strong

56   cold tongue forms along the southern California coast, and the surface pressure

57   increases in response to the cooling. The thermal structure indicates indirect circulation,

58   suggesting the conversion of eddy kinetic energy to eddy potential energy, implying that

59   the eddy kinetic energy is supplied from the large-scale northwesterly flow, but is

60   dissipated by potential energy conversion and surface friction.

61   The importance of diurnal variation and the lack of large-scale forcing are new

62   findings which are in sharp contrast to prior studies. These differences are due to the

63   inclusion of many short-lived eddy events detected in our study which have not been

64   included in other studies, aided by the hourly 10-km model output spanning 61 years.

65   Using extended historical analysis, the long-term behavior of the CE events is

66   examined. Although regression of the CE index onto surface temperature showed a La

67   Nina-like pattern, all the correlations were found to be insignificant. The correlations

68   with PDO, AO and ENSO indexes were also found to be small, suggesting weak

69   climate-scale control of CE activity by those modes.

70

71 **1. Introduction**

72   The Catalina Eddy (CE) is a formation of meso-scale cyclonic circulation along

73 the coast of southern California within a large-scale northeasterly flow associated with a

74 semi-permanent subtropical high pressure system. CE develops when the regular

75 northwesterly flow changes direction and turns to southerly flow along the California

76 coast, producing cyclonic circulation over the Southern California Bight. It has several

77 important meteorological implications including change in visibility, occasional

78 production of light rain, variation in pollution, and lowering of coastal temperature.

79 Some of the cloudy weather often associated with the CE is locally called "May gray"

80 or "June gloom" depending on the month of occurrence. It is an interesting

81 phenomenon dynamically, since the winds adjacent to the coast are directed towards the

82 opposite direction expected from the general large-scale pressure gradient.

83   A CE event is considered to be an instance of the occurrence of Coastally

84 Trapped Disturbances (Mass and Steenburgh, 2000), but it is differentiated by its

85 location and stationary nature in the Southern California Bight region. There are large

86 numbers of CE studies. Bosart (1983) was the first to conduct a comprehensive

87 synoptic analysis of an event using a limited number of observations, and to present the

88 evolution and possible mechanisms of its formation. His results were later confirmed

89   by several other studies, ranging from observational (Mass and Albright, 1989, hereafter

90   referred to as MA) to modeling (Davis, et al, 2000). The general nature of the CE and

91   its formation/decaying mechanisms is thought to be well understood. Currently well-

92   recognized characteristics of the CE are:

93   1) Eddies have a horizontal scale of about 100 km and vertical extension of

94   100-150 hPa.

95   2) The increase of stratus along the coast.

96   3) Most eddies are stationary, but occasionally have northward movement.

97   4) Events are associated with higher inversion layer.

98   5) The associated clouds extend well into land and prolong the cloudy state

99   into afternoon.

100   The dynamical mechanism is thought to be due to the effect of mountain wave

101   over the San Rafael Mountains which creates low pressure at the lee side of the

102   mountains, along the coast of the Santa Barbara Channel, triggering the eddy (Bosart

103   1983). The large-scale forcing that creates ridging along the coast, together with the lee

104   side trough, generates a south-to-north pressure gradient that forces the southerly flow

105   along the coast (Bosart, 1983; Clark, 1994; Reason and Steyn, 1992; Ulrickson et al

106   1995). Other theories suggest Kelvin wave formation due to the existence of a strong

107   inversion layer and a sudden change in large-scale forcing (Clark, 1994; Dormán, 1985;

108   Eddington et al, 1992). The decaying stage is associated with the weakening of the

109   along-shore pressure gradient due to the passage of synoptic disturbances. In all these

110   studies, large-scale forcing is one of the key controlling mechanisms.

111   Previous studies of CE events are mostly based on a few long lasting and well

112   pronounced cases. Bosart (1983) chose a case based on the satellite visible pictures.

113   Clark and Dembek (1991) utilized a special observation network together with GOES

114   satellite images. Wakimoto's (1987) analysis is based on a special observation program

115   with subjective surface pressure analysis. Kessler and Douglas (1991) also utilized a

116   special observation network to study the Eddy. Davis et al, (2000), Ueyoshi and Roads

117   (1993) and Ulrickson et al (1995) chose the Eddy case from MA.   Only MA utilized a

118   large number of cases (about 50) to study the general characteristics of the CE events. It

119   should be emphasized that in their study, CE is not defined by identifying the cyclonic

120   circulation, but by a proxy, namely by the long-lasting strong southerly flow at San

121   Diego. Thompson et al (1997) utilized a typical large-scale pattern favorable for Eddy

122   formation from the MA study and identified the Eddy by actually running the numerical

123   simulation.

124   All these studies placed a strong constraint on the time scale of the eddy by

125   restricting its lifetime to 18 hours or longer, which may alter the general nature of the

126   CE if the eddy's lifetime is shorter. The application of proxy data used by MA may

127   contaminate the CE cases when a sub-synoptic disturbance unrelated to CE passes

128   through, or the coastal wind reversal develops without forming eddy circulation.

129   Simply identifying the Eddy from one-time satellite observation can be dangerous in

130   some instances. In addition, strong diurnal variation of the wind system along the coast

131   noted by MA may greatly modulate the Eddy, whose analysis requires hard-to-obtain

132   high temporal resolution observations.   In fact, many of the uncertainties of CE studies

133   are due to the lack of proper high temporal and spatial observation networks.

134   Consequently, the characteristics and dynamical mechanisms of CE still have room for

135   further study.

136   The purpose of this paper is to utilize 57-year California Reanalysis

137   Downscaling at 10 km (CaRD10) resolution (Kanamitsu and Kanamaru, 2007;

138   Kanamaru & Kanamitsu, 2007), recently extended to 61 years, to examine the general

139   character of the Catalina Eddy and its possible formation mechanisms.   This dataset

140   consists of hourly 10 km horizontal resolution outputs for the years 1948 to 2008, ideal

141   for the detection of Catalina Eddy events. The output provides various diagnostic fields,

142   suitable for dynamical and thermodynamical analysis of the eddy. The hourly output

143   also allows us to study the interaction between diurnal variation and Eddy activity. In

144   addition, the 61 years of analysis enable us to examine the interannual and long-term

145   trend of the Eddy activity.

146   This paper will be structured in the following manner. Section 2 briefly

147   describes CaRD10, Section 3 provides the eddy detection method and its verification,

148   Section 4 is devoted to the climatological characteristics of the Eddy, Section 5

149   examines the dynamical and thermodynamical features together with the formation

150   mechanism, and Section 6 describes long-term climatological features. The summary

151   and conclusions are given in Section 7.

152   **2. CaRD10**

153   CaRD10 is a dynamical downscaling of NCEP/NCAR reanalysis. The 6 hourly

154   200 km resolution Reanalysis is dynamically downscaled to hourly 10 km resolution

155   using the RSM (Regional Spectral Model, Juang and Kanamitsu, 1994).   Spectral

156   nudging is applied to wind whose scale is greater than 1000 km (Kanamitsu et al, 2010).

157   Area average temperature at each level is also adjusted towards Reanalysis. Note that

158   the downscaling utilizes large-scale forcing only from reanalysis and no local

159   observations are used.   Therefore, small-scale details, including meso-scale CE

160   circulations, are simply generated by the regional dynamics/physics of the model. The

161 RSM is a state-of-the-art atmospheric model with a comprehensive physics package.

162 Details of the RSM are found in Kanamitsu and Kanamaru (2007). The CaRD10

163 product has been verified against station observations and compared with North

164 American Regional Reanalysis (NARR, Mesinger et al., 2010), which utilized a data

165 assimilation system with local surface observations. It was shown that over the ocean,

166 CaRD10 near-surface wind analysis is as good as NARR, and it is superior over land

167 (Kanamaru and Kanamitsu, 2007) due mainly to the higher resolution of RSM (10 km

168 vs. 35 km for NARR).

169 **3. Eddy detection method**

170   CE events are defined as near-surface cyclonic circulation localized within the

171 bight of southern California and centered near Santa Catalina Island (see Figure 1 for

172 characteristic geographical features and major city locations over the area).   A

173 composite of the wind field 10 m above the surface was used for CE events detection.

174 The canonical eddy was obtained by averaging fifteen typical events detected in both

175 CaRD10 and MA. Figure 2 shows the 15 events composite maps of the near-surface

176 relative vorticity and streamlines. The structure of the eddy is characterized by

177 elongated positive vorticity emanating from Point of Conception, with a narrow band of

178 negative vorticity along the coast.

179   The spatial pattern correlations between the composite near-surface wind

180 directions and the hourly CaRD10 10 m wind directions were then calculated within the

181 rectangular area 32°N – 34°N, 121° W- 117° W (area shown in Figure 2). Since we

182 were interested only in the wind directions and not in the amplitudes, we used the

183 Pearson product-moment correlation coefficient. We considered that a CE event

184 occurred if the correlation coefficient was larger than a certain threshold. We

185 intentionally did not segregate events by season, duration, or time of occurrence.

186 **3.1. Sensitivity to the value of the threshold**

187   Experiments to determine the sensitivity of the features of the detected eddy to

188 the correlation coefficient threshold were conducted. The composites of the initial

189 stages of the CE events for different threshold values were examined. Figure 3 displays

190 vorticity and wind detected with the threshold from 0.9 (upper left panel) to 0.1 (lower

191 right panel) in 0.1 increments. From subjective evaluation, a correlation of 0.7 which

192 presents significant cyclonic rotation over the Southern California bight, is selected for

193 the CE detection criteria. We note that most of the results obtained in this paper are not

194 very sensitive to the choice of this criterion, except for the distribution of the duration of

195 very short lasting events of about one hour.

196 **3.2. Validation**

197   Validations against other independent data sets were conducted.   First we

198   compared CE events detected with our method to those from MA for May –September

199   1968-1982. MA used the sustained southerly surface wind of at least 1. m/s for at least

200   18 continuous hours at a San Diego land station as a proxy for the CE event. Table 1

201   shows hit and miss occurrences for this time period. The success rate is 80% indicating

202   that the dynamical downscaling is capable of reproducing CE events without injection

203   of local observations and confirms that our method is a reasonable tool for CE event

204   detection. Ueyoshi and Roads (1993), Skamarock et al. (2002) and Davis et al. (2000)

205   have demonstrated the capability of numerical models to reproduce typical CE. The

206   events that were missed by our detection methods were subjectively analyzed. Some of

207   them   were   merely   coastally   trapped   disturbances   without   a   notable   rotational

208   component, and some others were baroclinic disturbances that happened to satisfy MA's

209   condition at San Diego, suggesting that our true CE detection rate is even higher.

210   In   order   to   further   validate   our   CE   detection   method,   we   used   three

211   independent observations. First we applied the 4 km resolution GOES satellite imagery

212   data available for the period December 1995 to September 2008. Some of the cases are

213   well-validated by anomalous low cloud occurrence, but validation of all the eddy cases

214   was very difficult due to difficulties in recognizing eddy patterns from the imagery.

215   Secondly, we validated against the North American Mesoscale (NAM) Analysis data,

216   which consist of 12 km horizontal resolution data assimilation utilizing conventional

217   surface and buoy observations. The data is available in six hour intervals (00, 06, 12 and

218   18 UTC). 75% of CaRD10 CE events were also found in NAM. Thirdly, as an attempt

219   to validate the accuracy of the diurnal component of the CE detected in CaRD10, we

220   utilized hourly National Data Buoy Center data. For this validation, the duration of the

221   southerly wind events at San Diego Scripps Pier, used as a proxy of eddy formation by

222   MA, is examined. Figure 4 shows PDFs of the duration of such events by hour. In the

223   Scripps Pier data, the short-lived southerly wind occurred nearly twice as often as in the

224   CaRD10 data, suggesting that the CaRD10 may even be underestimating the existence

225   of very short-lived Eddy events, but the relation between the Scripps Pier data and the

226   existence of CE is not certain for these very short-lived southerly events.

227   With these extensive comparisons to independent datasets, we are confident

228   that the CaRD10 dynamically downscaled analysis detects Catalina Eddy with

229   reasonable accuracy.

230   **4. Climatological characteristics**

231   Analysis of 61 years of CaRD10 Catalina Eddy events, summarized in Figure

232   5, revealed that the CEs appear most frequently in the May-June period when the eddies

233   are strongest in terms of vorticity (Fig. 5a and b). The maximum intensity eddy in terms

234   of vorticity is observed at 16 UTC, 8:00 a.m. local Pacific Standard time (Fig. 5d). The

235   average duration of the mature stage (defined as the wind direction correlation with

236   canonical CE greater than 0.7) is about 4.4 hours (Fig 5c), but a large number of eddies

237   have a mature stage shorter than 1 hour (Fig. 5d). When the threshold correlation for

238   CE detection is lowered to 0.5, the number of eddies with a mature stage shorter than 1

239   hour increases.   Mature stage eddies lasting longer than 18 hours are very rare,

240   accounting for only a small percent of all the CEs detected. This is in sharp contrast to

241   previous works which are based on long-lasting and relatively rare very strong cases.

242   The relatively short average mature stage duration obtained in this study is somewhat

243   surprising.   Considering the underestimation of San Diego offshore southerly wind

244   events in CaRD10 compared with offshore buoy observation mentioned in Section 3.1,

245   the actual number of eddies may be even larger and their mature stage duration may be

246   even shorter.

247   The reason for the short mature stage duration is further examined by looking

248   at an hourly evolution of eddies. Figure 6 shows an example of a short-lived eddy case

249   observed during April 30 and May 2, 2000. The eddy appeared consecutively for three

250   days at a particular local time, but it tended to weaken and disappear during 21 UTC to

251 03 UTC (1:00 p.m. to 7:00 p.m. local time). The hourly resolution CaRD10 output

252 makes it possible to identify this strong diurnal variability in the eddy activity. We

253 believe this intermittency is a reason for the relatively short mature stage duration

254 obtained in this analysis. The duration of mature stage CE can be aliased to a longer

255 lasting eddy if any of the observations are missing during 21 UTC to 03 UTC, or if the

256 relation between southerly wind in San Diego and Eddy formation breaks down. One

257 may notice that the slight southerly component at the San Diego location (indicated by a

258 white circle in the top left panel) is observed even when the CE dissipates, suggesting

259 the possible breakdown of the relationship between winds at San Diego and CE, leading

260 to the erroneous overestimation of eddy lifetime. Unfortunately, the evolution of CE

261 shown in Fig. 6 could not be verified against observations due to missing hourly station

262 data over the bight area. In order to further strengthen our argument, we show another

263 case with hourly station observation in Figure 7 (September 17-18, 2007). We see a

264 dissipating eddy at 01 UTC which is supported by several observations, but the eddy

265 reappears after 8 hours at 09 UTC.

266   The hour of the establishment of the mature stage and the maximum intensity

267 during the day are shown in Fig. 5e.   In terms of the pattern correlation, the maximum

268 number of cyclonic circulations are observed at around 08-13 UTC (0:00 a.m. to 4:00

269   a.m. local time).   The eddy reaches its maximum intensity, as measured by positive

270   vorticity amplitude, during 13-19 UTC (5 a.m. to 11a.m. local time). A more detailed

271   timeline of the CE life cycle will be discussed later.

272   Figure 8 shows the large-scale anomaly patterns at 500 and 850 hPa associated

273   with the mature phase of the CE. At 500 hPa, the typical pattern is the low anomaly to

274   the north and elongated strong positive anomaly extending from the western extra-

275   tropical Pacific towards inland, passing just north of Point Conception (Fig. 8a).   The

276   full field map shows that the geopotential height field is nearly zonal with a very weak

277   horizontally tilted trough oriented from 33N, 114W to 18N, 120W (Fig 8c), suggesting a

278   lack of strong large-scale cyclonic/anti-cyclonic forcing at least at this level. At 850

279   hPa, the negative anomaly extends southward over the land and a positive weak

280   anomaly over the western extra-tropics without clear ridging extension (Fig.8b) can be

281   seen. The anomalous wind is generally north-north-westerly over the entire domain

282   except over the southwestern portion.   The full field map shows generally high

283   geopotential over the southeastern portion of the Pacific and low geopotential over the

284   continent, producing a pressure increase from southwest to northeast (Fig. 8d).   The

285   anomaly patterns at these two levels agree well with MA, but the amplitudes of the

286   ridges and troughs are much weaker.   The anomaly at 850 hPa enforces wind flowing

287   across the San Rafael Mountains, which is a key to the generation of the CE, as detailed

288   in the following sections (also noted by all the previous studies).

289   Figure 9 (left panel) describes the main characteristics of the mature phase of

290   the CE anomaly composite. Note that the anomalies in this paper are computed by

291   subtracting the 57-year mean of the field at a fixed local time to eliminate the

292   climatological diurnal variability appearing in the anomaly field. The composite near-

293   surface structure of the eddy is characterized by elongated positive vorticity emanating

294   from the Point of Conception, with a narrow band of negative vorticity along the coast.

295   The location of the center of the positive vorticity coincides with the circulation center,

296   which is not the case in the composite eddy used for detection. There are two vorticity

297   maxima and minima roughly aligning east-west in pairs. The negative vorticity center

298   located just over the Santa Monica Mountains is more conspicuous, stronger and is

299   located to the south of the eddy composite used for detection (Figure 2). This negative

300   vorticity center enhances the onshore flow to the north, separating the positive vorticity

301   streak from Point of Conception into two maxima. It appears to be more distinct at

302   upper levels, clearly seen as anticyclonic eddy at 925 hPa (Fig. 9 right panel). This

303   anticyclonic circulation seems to be forced by the underlying geographical features of

304   the Santa Monica Mountains.

305   There is a surface convergence associated with the eddy (Fig. 10 left panel)

306   with the accompanying large areas of divergence to the east and northeast of the eddy.

307   Away from the eddy areas, areas of divergence anomaly are evident over the upslope

308   wind areas of the northern San Rafael Mountains. A belt of convergence along the coast

309   of the Santa Barbara Channel to the north of Los Angeles is the result of accelerating

310   down-slope wind from the mountains (or the lee wave) colliding with the northerly flow

311   associated with the CE. Near-surface temperature is colder to the east of the eddy

312   center and warmer to the west (Fig. 10, right panel), resembling the middle-latitude low

313   but with a completely opposite temperature structure, i.e., warm to the west and cold to

314   the east. Compared to the divergence field (Fig. 10, left) we see that the relatively cold

315   air is converging and rising and warm air is diverging and sinking, suggesting the

316   conversion from kinetic energy to potential energy.

317   There is a very warm anomaly area to the north of the CE, over the San Rafael

318   Mountains along the Santa Barbara Channel coastline. The warm area is located over

319   the mountain slopes with down-slope winds, suggesting the effect of adiabatic warming

320   caused by large-scale forcing consistent with the 850 hPa height and wind anomaly

321   (Fig. 8). The warm down-slope wind area along the Santa Barbara Channel is

322   indicative of the mountain effect, first described by (Bosart, 1983) and later by many

323   others (Davis et al 2000; Skamarock et al 2002; Thompson et al 1997 ; Ulrickson et al

324   1995).   Interestingly, southwest of Los Angeles in the San Diego Coastal Range (see

325   Fig. 1), a warm area is associated with the up-slope wind, suggesting the diabatic

326   heating control.   This is a good demonstration of different thermal and dynamical

327   mechanisms working in a small scale. The moisture anomalies show increased specific

328   humidity as well as increased relative humidity to the east of the Eddy, extending inland

329   (Fig. 11), indicating the intrusion of stratus farther into land.   Latent heat flux is

330   decreased by more than 25 W/m**2 along the vorticity center due to weaker winds (see

331   Fig 12).   This reduced evaporation may compensate the reduction in incoming solar

332   radiation due to increased cloudiness. Further study utilizing a regional coupled model

333   could yield interesting CE air-sea interaction.   The SLP pattern exhibits the low

334   pressure located at the center of the eddy circulation (Fig. 13) with a trough extending to

335   the Santa Barbara Channel coast.   The ridging along the southern California coast is

336   distinct. Clark (1994), Reason and Steyn (1992), Eddington et al (1992), Thompson et

337   al (1997), and Ulrickson et al (1995) consider this feature to be the key factor of the

338   initiation of the eddy.

339   The vertical cross sections of the eddy (Figure 14 right panel) show that the

340   positive vorticity anomaly extends to about 900 hPa. The negative vorticity to the east

341 of the eddy extends further into the 800 hPa level. The warm anomaly at around 950

342 hPa just above and west of the center is a very distinct thermal structure. A cold

343 anomaly exists to the east of the eddy center. Both warm and cold temperature

344 anomalies extend to 800 hPa. The thermal structure is not symmetric with respect to the

345 center of the eddy exhibiting an eastward vertical tilt, contrary to the middle latitude

346 disturbances.

347 **5. Development mechanism**

348 In this section the initiation phase is studied. In order to eliminate overlapping

349 events, we repeated analysis only for CE events separated by 24 or more hours from

350 other events. For this reason we avoided examining the decay phase, since 24 hour

351 separation can be contaminated by the initiation phase of the following CE event.

352 Figure 15 shows Z500 and Z850 fields at 21, 18, 15, 12, 9, 6, 3, and 1 hours before the

353 detection of a CE event, i.e. the start of the mature stage. Interestingly, there is very

354 little change in the 500 hPa height field prior to the CE formation, except for a slight

355 increase in the meridional gradient as the CE forms. This lack of large-scale

356 cyclonic/anticyclonic forcing during the initiation phase is in sharp contrast to MA and

357 other studies, in which large-scale forcing is found to be essential for initiation. This

358 result is due to the inclusion of many weak CE events in the current study, but

359   demonstrates that a weak CE can form without large-scale cyclonic/anti-cyclonic

360   forcing.

361   At 850 hPa, the situation is very much the same but an increase in the

362   intensity of the low over the Southeastern California/Southern Nevada region is noted,

363   which contributes to the increase of wind over the San Rafael Mountains by enforcing

364   the northerly flow. Figure 16 shows the near-surface anomaly wind vector and anomaly

365   vorticity during the initiation phase. At 21 hours before the mature stage, we see a small

366   area of positive vorticity located along the Santa Barbara Channel coast. The average

367   time of the full development phase is about 12 UTC (4:00 a.m. local time), thus, 21

368   hours before the full development stage corresponds to 15 UTC or 7:00 a.m. local time.

369   This vorticity maxima gradually intensifies and expands with time, and between 6 to 9

370   hours prior to the full development stage (03-06 UTC or 7:00 p.m.-10:00 p.m. local

371   time), there is explosive growth. At 3 hours prior (09 UTC or 1:00 a.m. local time), an

372   area of negative vorticity appears along the coast just to the east of the eddy which

373   eventually deforms the positive area to a shape resembling the sole of a shoe.   This

374   combined positive and negative vorticity region resulted in the formation of southerly

375   flow along the coast. The figure indicates that the origin of the eddy is at the coast of

376   the Santa Barbara Channel. Over this area, the streamline has cyclonic curvature due to

377   the flow around the San Rafael Mountains. The strong convergence of the streamline is

378   associated with the flow crossing over the mountain. A northeasterly anomaly over

379   ocean is also clearly observed. These features are commonly pointed out in all the

380   previous Catalina Eddy studies.

381   In order to examine the growth of the positive vorticity maxima, a vorticity

382   budget calculation is performed from composite maps. Figure 17 shows major vorticity

383   budget terms at various stages of initiation. We identified two controlling terms, a

384   stretching term and horizontal advection term. 21 hours before the mature phase (15

385   UTC or 7:00 a.m. local time), positive vorticity along the Santa Barbara Channel coast

386   is caused by the stretching term due to the convergence. The generated vorticity is

387   advected southeast by the quasi-stationary northwesterly flow. A very similar

388   description of the event is presented by Davis et al (2000). The local time change of

389   vorticity is relatively small, indicating that the generated and advected vorticity is

390   dissipated by the surface friction. At 12 hours prior to the full development phase (00

391   UTC or 4:00 p.m. local time), vorticity generation by the stretching term extends south

392   along the coast. Horizontal advection further stretches the area of the positive vorticity

393   anomaly southward along the coast. Thus, the major source of vorticity is still located

394   along the coast of the Santa Barbara Channel. 6 hours prior to CE detection (06 UTC

395 or 10:00 p.m. local time), generation of positive vorticity due to the stretching term

396 suddenly spreads all along the coast. This hour is found to correspond to the time when

397 the diurnal variation of wind causes the wind to turn from onshore to offshore flow

398 along the entire west coast, generating a convergence line along the coast. The diurnal

399 evolution of the wind is shown in Figure 18. The blue area (negative zonal wind, i.e.

400 easterly wind) develops all along the coast at 6 hours prior to the full development

401 stage, which continues for at least the next 6 hours. Apparently, this transition causes

402 strong convergence just a few kilometers offshore of the Bight. The positive

403 contribution of the stretching term is widespread over the entire Southern California

404 coast. Small (1999) noted the effect of coastal convergence on the Catalina Eddy.

405 Previous studies indicate that the effect of sea breeze is unimportant (Skamarock et al,

406 2002), or important but not crucial for eddy initiation (Davis et al, 2000; Ulrickson et al,

407 1995), which contradicts our results. Very close to the coast, horizontal advection by

408 the offshore flow creates an area of negative vorticity which intensifies as it approaches

409 the full development phase.

410 We comment here that the disappearance of the CE during 21 UTC and 03

411 UTC presented in Section 4 (Figs. 5 and 6) corresponds to the diurnal phase of onshore

412 flow receding to offshore flow, when the convergence is rapidly taken over by

413 divergence, weakening and dissipating the eddy. Thus diurnal variation also seems to be

414 crucial to the decay phase.

415   The evolution of the thermal structure during the initial stage of the CE is

416 shown in Figure 19. The anomaly near-surface air temperature evolution indicates

417 expansion of the warm area from the north, again emanating from the coast of the Santa

418 Barbara Channel. It gradually expands southeastwards for the period of 20 to 11 hours

419 prior to the full development stage due to advection by northwesterly flow over ocean.

420 This is similar to the evolution of vorticity, but directing towards the west side of the

421 eddy rather than to the center of the vortex since the interaction between temperature

422 and wind is weaker than that between vorticity and wind. The warm air to the east of

423 the Eddy center is a very clear signal of full development eddy from the cross section

424 map, already shown in Figure 14. About 10 hours prior to the full development stage, a

425 strong cold anomaly develops at the southwestern corner of the domain and extends

426 towards the east side of the eddy along the coast, reaching its maximum intensity at the

427 full development stage. The formation of a cold tongue along the coast is also noted in

428 the numerical simulation by Ulrickson et al (1995), who explained the cause as the

429 result of blocking of westerly and northwesterly wind over the bight combined with the

430 general warming to the west of the eddy. Corresponding to this cooling, a surface

431   pressure ridge forms along the southern California coast. This sequence of events

432   suggests that the north to south pressure gradient is not the mechanism of the Eddy

433   formation, but is rather its consequence. Except for the cold anomaly along the coast to

434   the south of the eddy, the temperature field more or less goes back to the stage 20 hours

435   prior.

436   Full field maps revealed (not shown) an intrusion of warm air from the south

437   towards the Eddy for the 20 to 11 hours prior to the full development stage but this

438   warm air intrusion considerably weakens during the last 10 hours, creating the strong

439   cold anomaly along the coast to the south of the eddy.   By the time the full

440   development stage is reached, the warm near-surface temperature anomaly along the

441   coast of the Santa Barbara Channel weakens.

442   In summary, from the vorticity budget view, the CE is initiated by the

443   stretching term due to the convergence caused by the flow across and around the San

444   Rafael Mountains. The generated vorticity is advected southward by the large-scale

445   wind associated with the subtropical high and dissipates due to surface friction at the

446   very early stages. As the generation by convergence continues, about 6 hours before the

447   full development phase, sudden generation of vorticity due to stretching occurs along

448   the entire west coast. This is caused by the convergence due to the turning of offshore

449   to onshore flow associated with diurnal variation.   The vorticity at the CE center

450   increases due to advection from the north and generation along the coast. At this time,

451   the CE takes its characteristic shape. At the same time, very close to the coastline,

452   negative generation of vorticity takes place due mainly to horizontal advection of

453   negative vorticity from land. The maximum negative vorticity center appears over the

454   Santa Monica Mountains.   This negative vorticity tends to separate the streak of

455   vorticity maximum emanating from the Santa Barbara Channel coast into two centers,

456   creating the shoe-sole shaped positive vorticity area.   The location of the negative

457   vorticity center positioned at the arch of the sole corresponds well to the Santa Monica

458   Mountain region, suggesting the effect of local small-scale topographic features.

459   The thermal structure indicates the importance of the down-slope wind along

460   the coast of the Santa Barbara Channel, causing a warm anomaly from adiabatic

461   heating.   This warm air is advected towards the southeast by the large-scale

462   northwesterly flow, causing a warm anomaly to the west of the Eddy center. At the fully

463   mature stage, a cold anomaly area appears along the southern coast.

464   Figure 20 shows hourly variation of surface pressure anomaly during the

465   initiation phase. 20 hours prior to the full development stage, there is a broad area of

466   low pressure anomaly over the entire Bight, which gradually deepens until 13 hours

467   prior when a small low pressure center area appears at the northern part of the Bight.

468   We do not see any low pressure area along the coast of the Santa Barbara Channel. The

469   low pressure center is located at the area with strong cyclonic curvature and diffluent

470   streamlines. The low pressure area gradually spreads towards southeast. At 10 hours

471   prior, an eddy center becomes recognizable at the western portion of the Santa Barbara

472   Channel coast. The low pressure area does not correspond to the center of the eddy, but

473   is located to the south/southeast, presenting the direction of movement of the eddy

474   center. The low pressure anomaly deepens further and at 6-5 hours prior, the eddy

475   center moves southeastwards and the pressure center starts to collocate with the eddy

476   center. The eddy further strengthens in harmony with the deepening of the low pressure

477   center afterwards. This wind-surface pressure anomaly relation shows that there is a

478   dominant ageostrophic component at the early stage, but the wind-pressure relationship

479   suggests that pressure is not driving the wind since the wind is not directing towards the

480   low pressure. Rather, the pressure anomaly seems to indicate the direction of the

481   movement of the Eddy at the early stage. This is in contrast with the Bosart (1983) and

482   MA interpretations of the development of the coastal southerly flow, which emphasize

483   the south to north pressure gradient as a driving mechanism. In our analysis, there is no

484   apparent north-south pressure gradient anomaly that drives the southerly flow, except at

485   the mature stage when the surface pressure ridge forms along the southern California

486   coast.

487   Figure 21 summarizes the life cycle of CE in relation to local time, together

488   with its relation to sea-breeze circulation. Some of the sample cases discussed in this

489   paper are also indicated. The rough estimate of the eddy's lifetime is 19 hours. The

490   eddy develops abruptly in the early evening and reaches the mature stage by midnight,

491   maintains its intensity until early morning, and then decays in the late afternoon, often

492   completely disappearing. The eddy is initiated when the phase of the sea-breeze turns

493   from onshore to offshore, and it decays during the opposite phase.

494   The current analysis suggests that the source of the eddy is the vorticity

495   generated by the convergence along the coast of the Santa Barbara Channel, which is

496   advected to the southeast to the Bight region, interacting with the vorticity generation

497   along the coast due to coastal convergence caused by the turning of the wind from

498   onshore to offshore sea breeze circulation. The event is vorticity driven rather than

499   pressure driven.

500   Looking at the thermal structure and convergence pattern, CE has a weak

501   baroclinic structure with temperature and vertical motion anomaly relationships,

502   convergence over a cold anomaly and divergence over a warm anomaly, indicating the

503 conversion of eddy kinetic energy to eddy potential energy. This implicitly indicates

504 that the CE's source of eddy kinetic energy comes from large-scale flow associated with

505 the sub-tropical high.

506 **7. Long-term climatological behavior**

507 The 57-year CaRD10 analysis and its extension allow for investigation of the

508 long term variability of CE events, which will be described in this section. In order to

509 characterize CE activity, we examined 4 indices; the monthly mean of vorticity

510 averaged over the CE region (shown in Figure 2), the monthly averages of wind

511 correlations used to detect the CE over the same region, the number of events per month

512 and the number of hours of observed CE per month. All these four indices are highly

513 correlated with each other with the lowest correlation being 0.64. Figure 22 compares

514 the first two CE indices, with a correlation coefficient of 0.75. The figure shows there is

515 very little long-term trend in the index, but some tendency of increase after 1990 may be

516 noted.

517 The temporal spectral analysis (Figure 23) of the vorticity averaged over the

518 area used above exhibits peaks corresponding to annual cycle, 10-year cycle and 26-day

519 cycle. The well known 26-day atmospheric cycle is the Yanai wave or the mixed

520 Rossby gravity wave in the tropics. In ocean, the 26-day cycle is observed in the

521 tropical Indian Ocean (Luyten, J. R., 1882) . The Tropical Instability Wave (Willett, et

522 al 2006) also has a periodicity of 20-40 days. However, all these waves are tropical in

523 origin, and their role in sub-tropical latitudes has not yet been identified. The 10-year

524 cycle probably corresponds to the PDO and is briefly discussed below.

525   The large-scale pattern associated with frequent eddy events is shown in Figure

526 24, presenting a strong ridging of anomaly over the west coast, corresponding to a more

527 zonal pattern. Near-surface temperature resembles a La Nina type of pattern, but

528 correlations do not exceed 0.3. These low correlation values are also found in the

529 individual correlations between low-frequency variability indices; ENSO correlation is

530 about -0.12, PDO correlation is about 0.05, and AO correlation is about 0.1 without lag,

531 and lower correlations with lags.   We tentatively concluded that the relations between

532 CE events and low frequency variabilities were not significant, and predicting the year-

533 to-year variation of CE events is very difficult.

534 **8. Summary, discussions and conclusions**

535   Climatological properties, dynamical and thermodynamical characteristics, and

536 long-term variabilities of the Catalina Eddy (CE) are examined from the 61 year

537 NCEP/NCAR Reanalysis downscaled to hourly 10 km resolution. The eddy is

538 identified as a cyclonic circulation confined in the Southern California Bight region.

539   Pattern correlation of wind vector from canonical CE is used to extract cases from the

540   downscaled analysis. Validation against published cases, the NCEP operational North

541   American Mesoscale Analysis, geostationary satellite images, surface observations and

542   buoy observations are made and confirm that the downscaled analysis accurately

543   reproduces Catalina Eddy events.

544   In this study, 4136 Eddy events with 16769 total mature stage hours were

545   identified during the 61 years. This is 3% of the total hours in 61 years. The detailed

546   dynamical and thermodynamical features of the Eddy were examined by compositing

547   these cases during the initiation and mature phases.

548   The large-scale features associated with the Eddy formation comprise a zonal

549   flow at 500 hPa and a horizontally tilted trough oriented from Northeast to Southwest

550   over the continent with low pressure over the Southern California/Southern Nevada area

551   at 850 hPa. The 850 hPa anomaly enforces the north-northeasterly flow over the San

552   Rafael Mountain region which has been identified as crucial for the initiation of the CE.

553   The favorable months for eddy formation are May and June, while the least

554   formation occurs during the winter months. The strongest CEs appear in June. We

555   found that the CE events are more frequent than the previous studies suggested, since

556   those studies chose only strong events. There are very few CEs that maintain their full

557   strength longer than 18 hours. The average duration of the full strength of a CE is of the

558   order of 4-5 hours. The relatively short duration of the mature stage in this analysis is

559   due to the use of hourly output, which shows interruption of the eddy by the onshore sea

560   breeze circulation during the late afternoon to early evening (around 4 p.m. to 6 p.m.)

561   local time. This interruption may have been missed in previous studies due to the lack

562   of surface observations necessary to directly identify the cyclonic circulation within the

563   Bight region.   The eddy intensity is strongest in the early morning at around 4:00 a.m.

564   local time.

565   The climatology of the mature phase of the eddy has the following

566   characteristics. A streak of vorticity emanating from the coast of the Santa Barbara

567   Channel is separated into two centers by a strong small-scale negative vorticity center

568   located over the Santa Monica Mountains.   The southern vorticity maximum

569   corresponds to the center of the Catalina Eddy.   The vorticity along the coast is

570   generally negative. The most prominent divergence feature is the strong divergence

571   over the northern slope of the San Rafael Mountains, paired with the strong

572   convergence over the southern slope. The region of positive vorticity streak and the

573   center of the eddy are associated with convergence. Other notable areas of divergence

574   are just off the coast of the Point of Conception and along the coast of Southern

575 California to the southeast of the eddy center. The temperature anomaly is warm to the

576 west and north of the eddy center, and cold to the west and southwest. There is an area

577 of very warm anomaly along the southern downslope region of the San Rafael

578 Mountains. The moisture anomaly shows an increase over the eastern part of the Eddy,

579 extending well into land, suggestive of the increase in cloudiness. Evaporation from the

580 surface is reduced at the center and to the east of eddy (but offshore) due to reduced

581 wind speed. Far offshore over the ocean not affected by eddy, evaporation is increased

582 due to increased wind speed, suggesting the possible importance of local air-sea

583 interaction. The surface pressure anomaly is in geostrophic balance with the near-

584 surface streamline, with minimum pressure coinciding with the eddy center. The

585 vertical extent of the eddy reaches only to 900 hPa as shown by previous researchers.

586 The thermal structure is warm to the west of the eddy and cold at the eddy center and to

587 the east. This thermal structure is in agreement with Ulrickson et al. (1995).

588 Temperature anomalies reach 800 hPa. The thermal structure is not symmetric,

589 suggesting an eastward tilt with height.

590 The analysis of the development stage of the eddy from composite maps shows

591 that the large-scale 500 hPa pattern does not change very much during the initiation,

592 except for the strengthening of the low located at southeastern California/southern

593   Nevada at 850 hPa. This geopotential height anomaly tends to enhance north-north

594   easterly flow across the San Rafael Mountains. The vorticity anomaly maximum is

595   located at the coast of the Santa Barbara Channel 20 hours prior to the fully developed

596   stage, slowly expanding along the coast, with explosive development at around 6 hours

597   prior to the mature stage. The vorticity budget indicated that the positive anomaly along

598   the Santa Barbara Channel coast is caused by the stretching term due to convergence

599   associated with the flow crossing over the San Rafael Mountains. This vorticity is

600   advected by large-scale northwesterly flow, but its magnitude is not large enough to

601   form a cyclonic circulation. About 6 hours prior to the start of the mature stage, there is

602   a strong generation of vorticity due to the stretching term all along the coast. This

603   stretching is caused by coastal convergence due to the phase change of the sea-breeze

604   circulation turning from onshore to offshore.   Ulrickson et al. (1995) noted that sea-

605   breeze circulation simply delays the occurrence of the Eddy. Davis et al (2000) noted

606   the importance of sea-breeze. Clark and Dembek (1991) noted strong interaction

607   between the sea-breeze and CE. However, none of these studies noted the critical

608   importance of the sea-breeze circulation in initiating the CE. Our analysis clearly

609   indicated that the phase change of the sea-breeze circulation is essential for eddy

610   formation and decay.

611   The difference in the importance of the sea-breeze can be traced back to the

612   difference in large-scale forcing between the current and previous studies.   Our

613   composite study indicated that the large-scale cyclonic/anticyclonic forcing was not

614   necessary for the eddy formation, while all the previous studies focused on the strong

615   eddy and concluded that the large-scale forcing is crucial.   Our study suggests that the

616   eddy will form without strong large-scale forcing, and in this situation, sea-breeze

617   becomes essential for the control of the evolution of the eddy.   Under strong large-scale

618   forcing, it is most likely that the effect of the sea-breeze circulation on eddy evolution

619   becomes secondary.

620   The near-surface thermal structure of the eddy evolves with a warm anomaly located

621   at the coast of the Santa Barbara Channel due to descending air along the San Rafael

622   Mountains forced by 850 hPa northeasterly winds. This warm air is advected southward

623   by the large-scale northwesterly flow.   Because of the weaker interaction between

624   temperature and wind field compared to that between vorticity and wind, the warm air

625   tends to spread over the area to the west of the eddy center, creating the warm anomaly

626   there. The downward motion due to the acceleration of northwesterly flow off the Point

627   of Conception may also contribute to the warming.   On the contrary, a cold anomaly

628   develops to the south and southeast of the eddy along the coast. This is probably due to

629 the increased vertical mixing caused by the rise of the top of the boundary layer

630 associated with the blocking of westerly winds by coastal mountains to the south of the

631 Eddy, combined with the general warming to the west of the eddy as suggested by

632 Ulrickson et al., (1995). These advection and vertical circulations are responsible for

633 the formation of the asymmetric thermal structure, as well as the north to south pressure

634 gradient along the coast. Although previous researchers argued that the eddy is initiated

635 by the surface pressure gradient along the coast, this study seems to indicate that the

636 pressure gradient is a result rather than a cause of the eddy formation, although as

637 argued by Ulrickson et al. (1995), the surface pressure gradient may act as a mechanism

638 to maintain the southerly flow along the coast after its formation. Our analysis seems to

639 indicate that the pressure pattern is determined by the eddy circulation and not the

640 reverse. This difference of the role of pressure gradient along the coast may also be

641 related to the differences in the strength of eddies studied in this and previous studies. It

642 is probable that in order to maintain the eddy for more than 18 hours, it is essential to

643 maintain a north to south pressure gradient by large-scale forcing.

644   The relation between divergence and temperature anomaly shows that the

645 relatively warm air is sinking to the west of the eddy and cold air is rising at the eddy's

646 center and to the east, implying energy conversion from eddy kinetic to eddy potential

647 energy. This conversion also implicitly indicates that the eddy kinetic energy is

648 generated from large-scale northeasterly flow, dissipated through kinetic to potential

649 energy conversion, together with the effect of surface friction.

650 Finally, the trend, inter-annual and inter-decadal variabilities of the CE

651 activities are examined. The temporal spectra of the vorticity averaged over the target

652 area shows peaks at 10 years, 1 year and 26 days. Regression of the CE index onto

653 monthly mean 500 hPa height indicates strong ridging in the anomaly over the west

654 coast, indicating the dominance of the zonal flow. Similar regression onto monthly

655 mean near-surface temperature showed a La Nina-like pattern with a warm anomaly

656 located over the west coast.   Unfortunately, the correlation between PDO, AO and

657 ENSO indices are found to be very low, of the order of 0.1-0.2, suggesting that the

658 large-scale control of CE is weak except for the local positive anomaly at 500 hPa.

659

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Figure Captions

Figure 1. Major topographical features and names of the main cities over the Southern California Bight region. The features referenced in the paper are shown in bold and underlined.

Figure 2. A composite of 15 events for the relative vorticity and streamlines. The composite near-surface structure of the eddy is characterized by elongated positive vorticity emanating from Point of Conception, with a narrow band of negative vorticity along the coast.

Figure 3. Composites of the CE events for correlation threshold of 0.9 (upper left panel) to 0.2 (lower right panel) in 0.1 intervals. Vorticity contour is shown in red.

Figure 4. PDF of the duration (hours) of CE events as measured by meridional velocity greater than 1.4 m/s for Scripps Station located at 32.867N 117.257W (blue) and CaRD10 wind in San Diego (magenta).

Figure 5. (a) Annual Cycle, (b) intensity, (c) average duration (d) PDF of duration (e) PDF of the Hours when CE was present/Diurnal cycle.

Figure 6. Evolution of 10-meter wind streamline and vorticity (shading) for 00 UTC April 30, to 21 UTC May 2, 2000. Note the complete disappearance of cyclonic circulation at 00 UTC (top row, first, third and 5th columns) and 03 UTC (same columns). Approximate location of San Diego is shown by white circle on the top left panel.

Figure 7. Another example of dissipating eddy but with surface observation plotted.

Figure 8. 500 and 850 hPa geopotential height anomaly composite of the mature phase of CE events (a) and (b) and 500 and 850 hPa geopotential height field composite of the mature phase of CE events (c) and (d).

Figure 9. Vorticity and wind anomalies at 10 m (left panel), and 925 hPa (right panel) for the composite of the mature phase of CE events.

758

759   Figure 10. Composite anomaly of divergence at 10 m (left) and temperature at 2 m.
760   The temperature map is overlaid by topography contours (in white).

761

762   Figure 11. Composite anomaly of specific humidity (left) and relative humidity (right).

763

764   Figure 12. Composite anomaly of latent heat flux (left) and wind speed.

765

766   Figure 13. Composite surface pressure.

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768   Figure 14. Composite cross section of temperature anomaly (left) and vorticity (right).
769   The center of the eddy is indicated by a triangle.

770

771   Figure 15. 500 hPa and 850 hPa geopotential height fields at 21, 18, 15, 12, 9, 6, 3, and
772   1 hours before the detection of CE event.

773

774   Figure 16. Anomaly streamline and vorticity at 21, 18, 15, 12, 9, 6, 3, and 1 hours
775   before the detection of CE event.

776

777   Figure 17. Vorticity budget at 10 m above ground at 21, 12, 6 and 1 hour prior to the
778   recognition phase. T3 is a stretching term, T1+T2 is the horizontal advection term and
779   T0 is a local time change or the storage term.

780

781   Figure 18. Near-surface wind vector at initial stages of the eddy formation. Shading is
782   for the east-west component of the wind highlighting the development of the offshore
783   wind (cold colors).

784

785   Figure 19. Composite evolution of near-surface air temperature anomaly during the
786   initial phase.

787

788   Figure 20. Hourly variation of surface pressure anomaly during the initiation phase of
789   the CE obtained from composite.

790

791   Figure 21. Illustration of the CE life cycle. Horizontal axis is time in local Pacific
792   Standard Time.
793   Figure 22. Comparison of two CE indices; I2.1 Monthly averages of vorticity averaged

794 over the CE region and I2.2 Monthly averages of winds correlations averaged over the

795 CE region. The correlation between the two indices was 0.75.

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797 Figure 23. Temporal spectra of vorticity averaged over the Southern California Bight

798 region (32°N – 34°N, 121° W- 117° W).

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800 Figure 24. Regression of the CE index onto NCEP/NCAR reanalysis monthly mean

801 500 hPa height (top) and near-surface temperature (bottom).

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Table 1: Hit and miss occurrences with respect to CE detected by MA for May-September 1968-1982

Init_Date	InitHour	Model InHour	End Date	EndHour	Model Dur
1968-05-01-21	2925	2918/2929	1968-05-02-21	2949	4HH;1HH
1968-05-27-09	3537	3536/3558	1968-05-29-03	3579	8HH;3HH
1968_06-23-15	4191	4183	1968-06-25-18	4242	11HH;
1968-06-28-18	4314	N/A	1968-06-30-21	4365	N/A
1969-06-15-15	3975	3976/3997	1969-06-17-21	4029	7HH/10HH
1969-06-23-21	4173	4163/4186	1969-06-26-18	4242	11HH/1HH
1970-06-12-00	3888	3878/3948	1970-06-14-00	3936	3hh/4hh
1970-06-28-12	4284	4288/4309	1970-06-29-21	4317	4HH/5HH
1970-09-11-18	6090	N/A	1970-09-13-06	6126	N/A
1971-05-05-21	2997	2993	1971-05-07-06	3030	3HH
1971-09-01-06	5838	5840	1971-09-03-06	5886	8HH
1972-07-14-06	4686	4688	1972-07-16-09	4737	2HH
1972-08-02-06	5142	5144/5154	1972-08-04-00	5184	8HH/2HH
1972-08-17-18	5514	5507	1972-08-19-15	5559	6HH
1972-08-23-06	5646	5646/5654/5670	1972-08-26-00	5712	7HH/2HH/9HH
1972-09-05-06	5958	N/A	1972-09-07-06	6006	N/A
1972-09-22-21	6381	N/A	1972-09-25-09	6441	N/A
1973-05-04-09	2961	2969	1973-05-05-09	2985	6HH
1973-06-13-12	3924	3930	1973-06-15-00	3960	8HH
1973-08-25-03	5667	5671/5675	1973-08-26-15	5703	2HH/9HH
1973-09-07-06	5982	5994/6005	1973-09-10-00	6048	2HH/3HH
1974-05-14-15	3207	3204/3224	1974-05-16-00	3240	10HH/12HH
1974-06-06-06	3750	3792/3796	1974-06-09-03	3819	2HH/5HH
1974-08-18-12	5508	5506/5530/5552	1974-08-21-06	5574	3hh/8HH/3HH
1975-07-12-06	4614	4618/4641/4669	1975-07-14-15	4671	9HH/3HH/1HH
1976-05-02-06	2934	2937/2947	1976-05-04-09	2985	7HH/1HH
1976-09-15-00	6192	6185/6202	1976-09-16-09	6225	3HH/2HH
1977-05-04-09	2961	2968/2984	1977-05-06-03	3003	3HH/11HH
1977-07-22-18	4866	4859/4878/	1977-07-24-06	4902	6HH/2HH
1977-08-17-15	5487	N/A	1977-08-18-09	5505	N/A
1977-08-26-09	5697	5699/5724	1977-08-28-12	5748	9HH/9HH
1978-06-26-09	4233	4231	1978-06-27-03	4251	13HH
1978-09-05-21	5949	N/A	1978-09-06-15	5967	N/A
1979-05-26-21	3501	3505/3541	1979-05-28-00	3528	1HH/6HH
1979-06-05-06	3726	3737/3752/3793	1979-06-08-06	3798	2HH/2HH/3HH
1979-06-17-09	4017	3993	1979-06-18-09	4041	10HH
1979-07-01-09	4353	4355	1979-07-02-21	4389	10HH
1979-07-12-15	4623	4625/4637/	1979-07-14-00	4656	2HH/2HH
1979-07-19-03	4779	4767	1979-07-21-00	4824	4HH
1980-05-07-06	3054	3063/3085	1980-05-08-06	3078	2HH/7HH
1980-06-30-12	4356	4355/4371	1980-07-02-00	4392	7HH/1HH
1980-08-02-12	5148	5147/5177/5199	1980-08-04-15	5199	6HH/1HH/3HH
1980-08-12-12	5388	5410	1980-08-13-21	5421	11HH
1980-09-07-09	6009	N/A	1980-09-08-21	6045	N/A
1981-06-08-12	3804	3808/3875	1981-06-11-12	3876	3HH/5HH
1981-06-17-18	4026	4019	1981-06-19-18	4074	8HH
1981-08-30-12	5796	N/A	1981-09-01-00	5832	N/A
1981-09-22-09	6345	N/A	1981-09-24-00	6384	N/A
1982-05-08-00	3048	N/A	1982-05-09-03	3075	N/A
1982-05-18-00	3288	3280	1982-05-19-09	3321	14HH

Table 2: CE occurrences based on CaRD10 and NARR surface data for May 2005. 75% of CaRD10 CE events were detected in NARR.

CaRD10	NAM
2005 05 04 07	YES
2005 05 05 03	YES
2005 05 05 20	YES
2005 05 14 11	YES
2005 05 16 07	YES
2005 05 20 09	NO
2005 05 23 03	YES
2005 05 24 06	YES
2005 05 27 14	NO
2005 05 28 04	YES
2005 05 29 14	YES
2005 05 31 13	NO
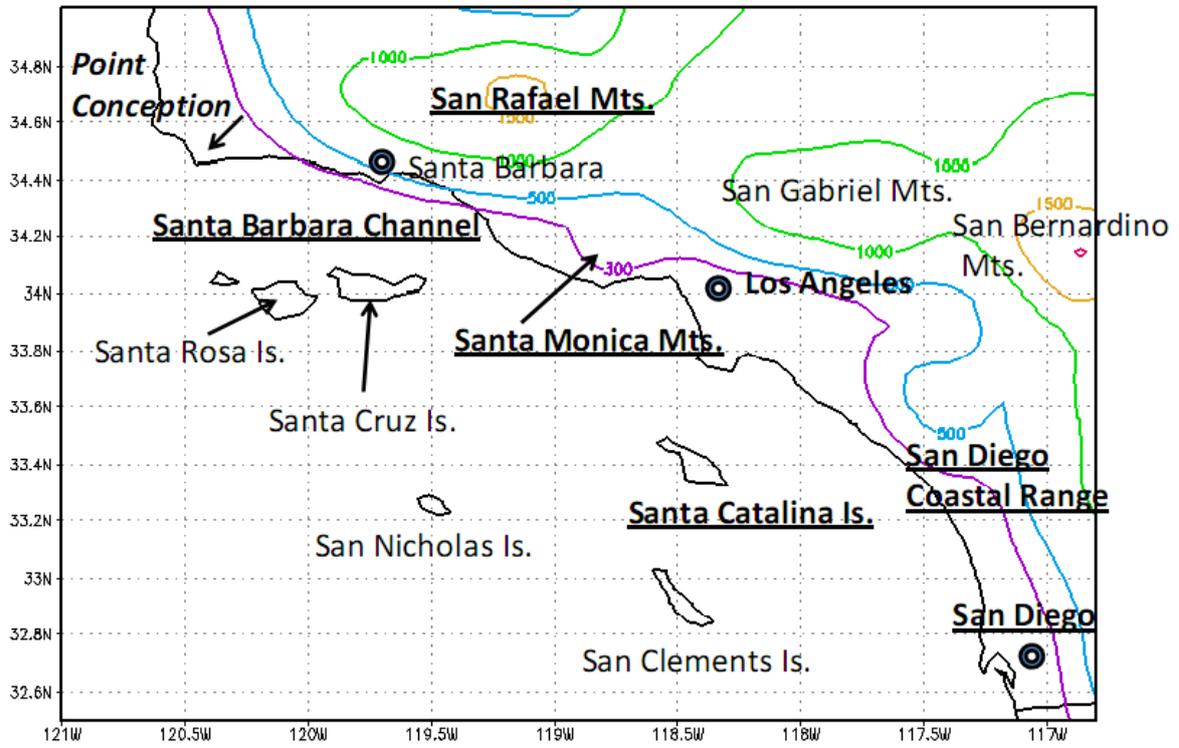
Figure 1. Major topographical features and names of the main cities over the Southern California Bight region. The features referenced in the paper are shown in bold and underlined.
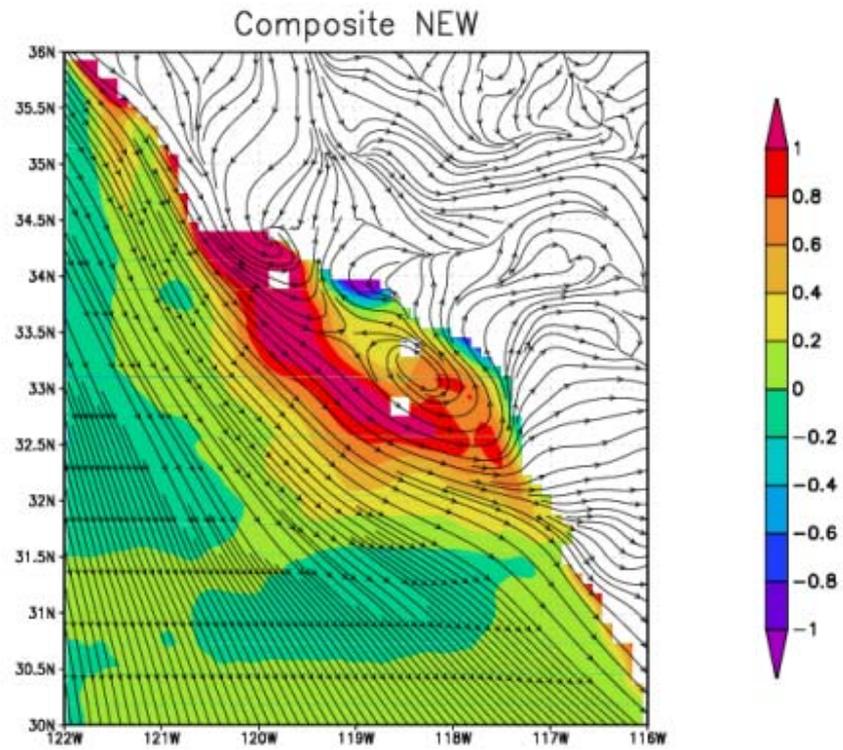
Figure 2. Composite of 15 events for the relative vorticity and streamlines. The composite near-surface structure of the eddy is characterized by elongated positive vorticity emanating from Point of Conception, with a narrow band of negative vorticity along the coast.

Figure 3. Composites of the CE events for correlation threshold of 0.9 (upper left panel) to 0.2 (lower right panel) in 0.1 intervals. Vorticity contour is shown in red.

Figure 4. PDF of the duration (hours) of CE events as measured by meridional velocity greater than 1.4 m/s for Scripps Station located at 32.867N 117.257W (blue) and CaRD10 wind in San Diego (magenta)

Figure 5. (a) Annual Cycle, (b) intensity, (c)average duration (d) PDF of duration (e) PDF of the Hours when CE was present/Diurnal cycle

Figure 6. Evolution of 10-meter wind streamline and vorticity (shading) for 00Z April 30, to 21Z May 2, 2000. Note the complete disappearance of cyclonic circulation at 00 Z (top row, first, third and 5th columns) and 03Z (same columns). Approximate location of San Diego is shown by white circle on the top left-most panel.
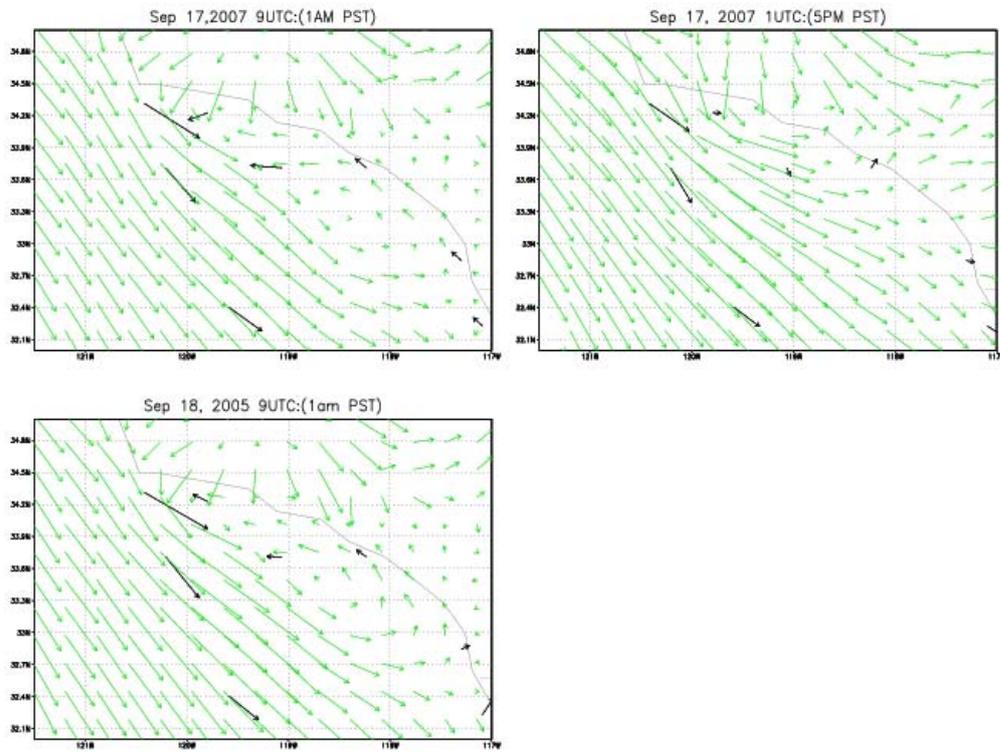
Figure 7. Another example of dissipating eddy but with surface observation plotted.
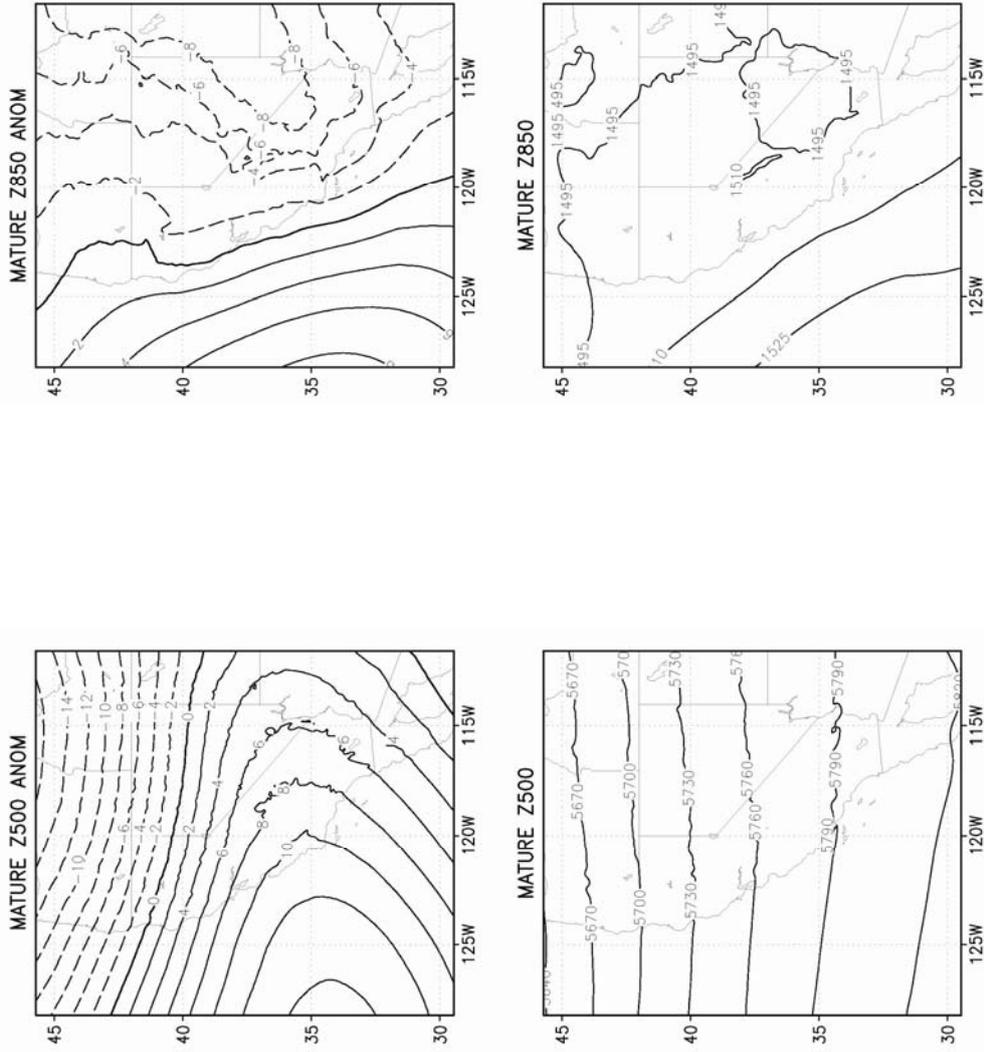
Figure 8. 500 and 850 hPa geopotential height anomaly (top two panels) and full field (bottome two panels) composite of mature phase of CE evens.
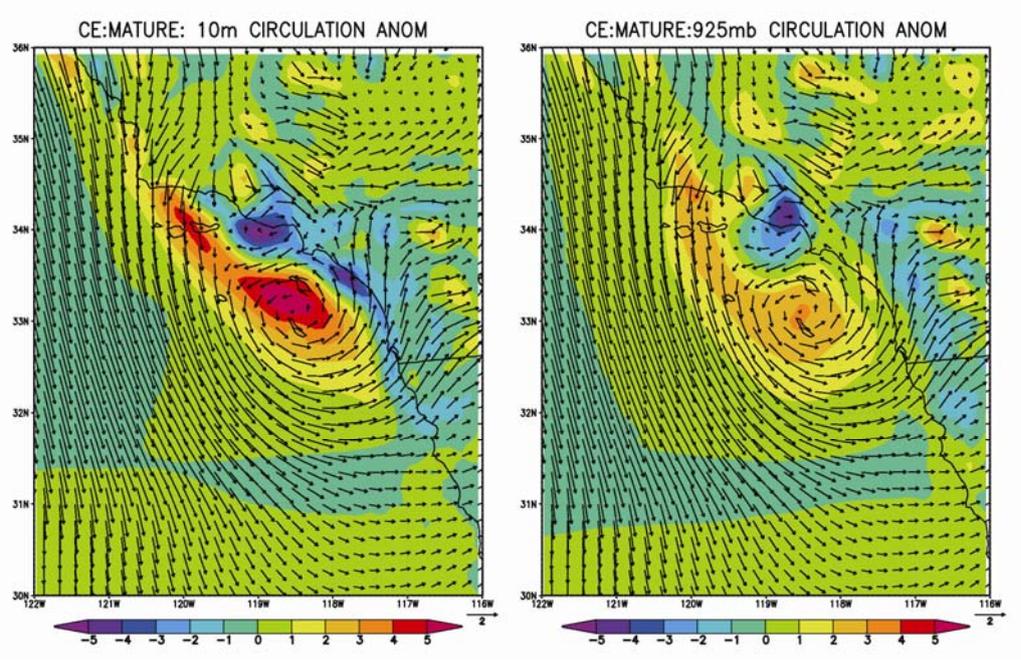
Figure 9. Vorticity and wind anomalies at 10m (left panel), and 925 hPa (right panel) for the composite of the mature phase of CE events.
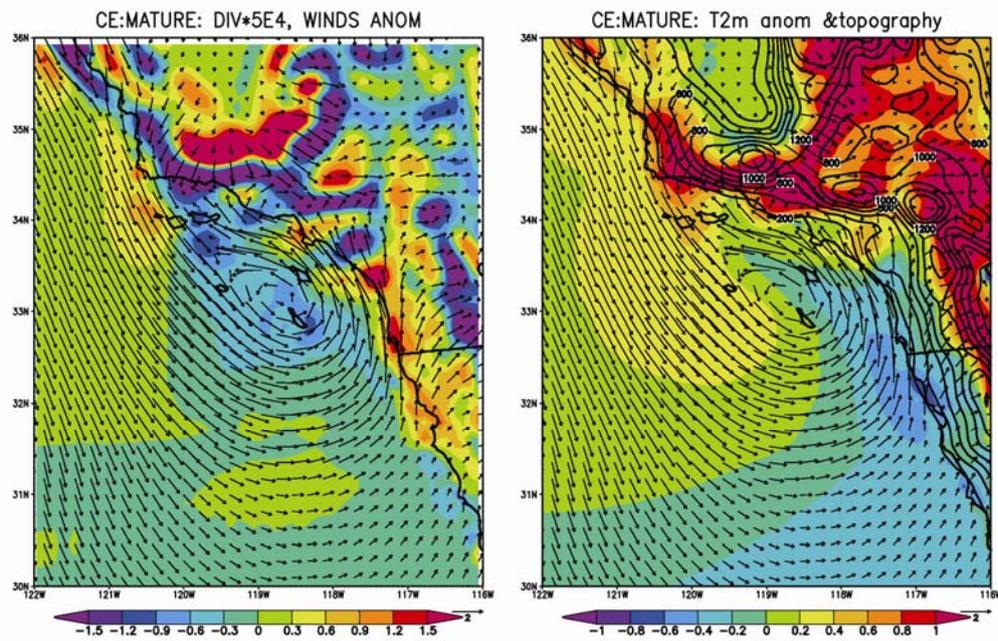
Figure 10. Composite anomaly of divergence at 10m (left) and temperature at 2m. The temperature map is overlaid by topography contours (in thick black lines).
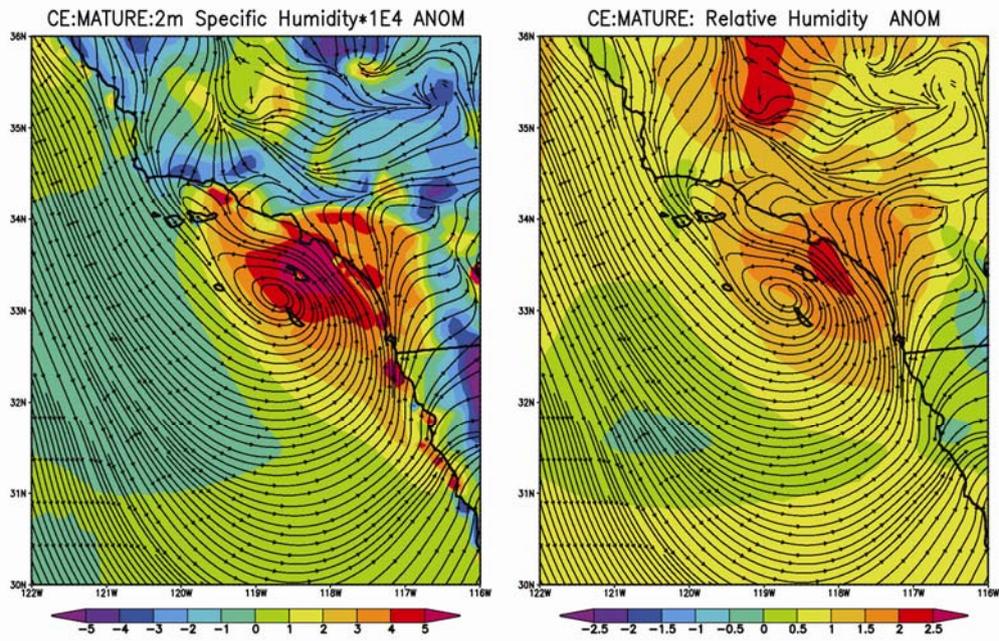
Figure 11. Composite anomaly of specific humidity (left) and relative humidity (right).
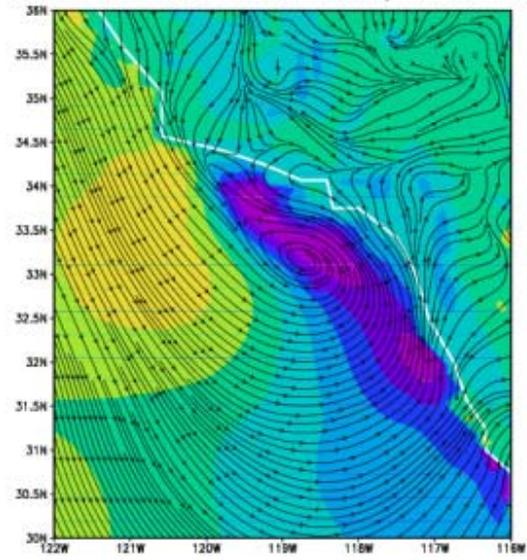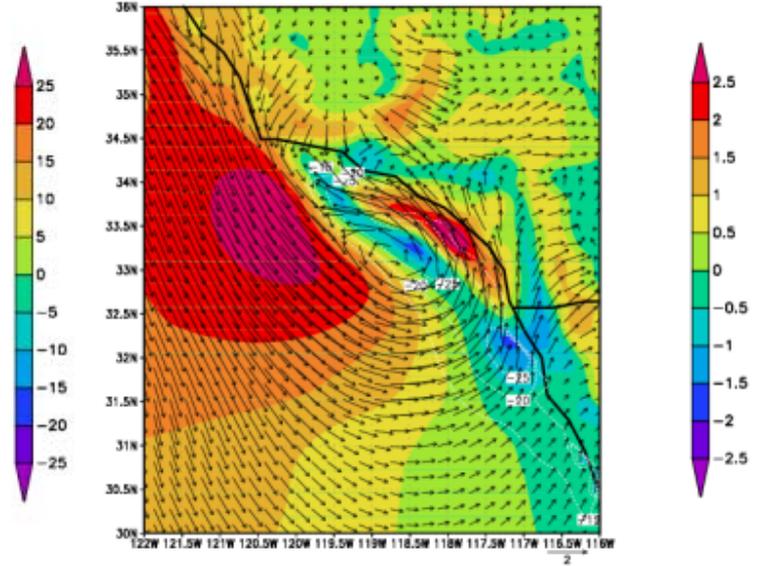
Figure 12. Composite anomaly of latent heat flux (left) and wind speed.
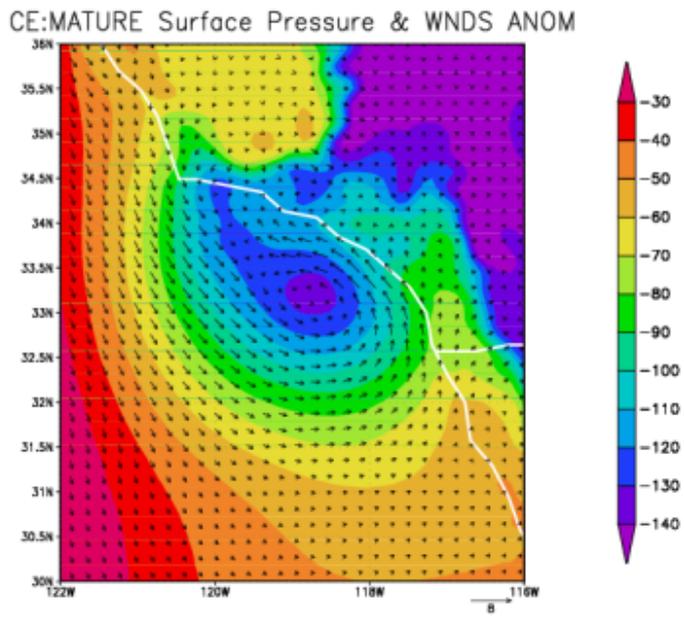
Figure 13. Composite surface pressure.
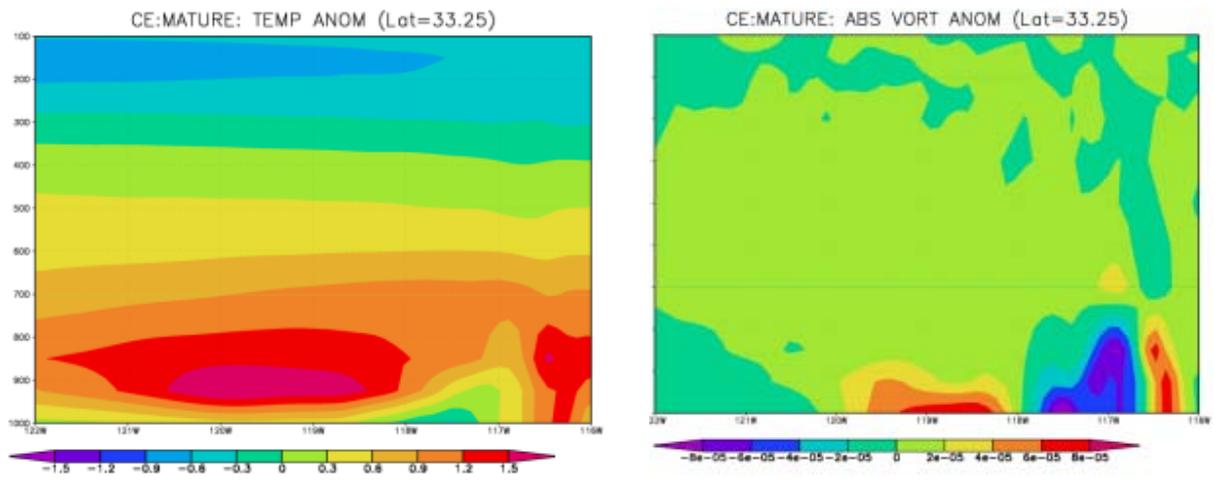
Figure 14. Composite cross section of temperature anomaly (left) and vorticity (right) along east west crossing the eddy center. The center of the eddy is indicated by triangle.
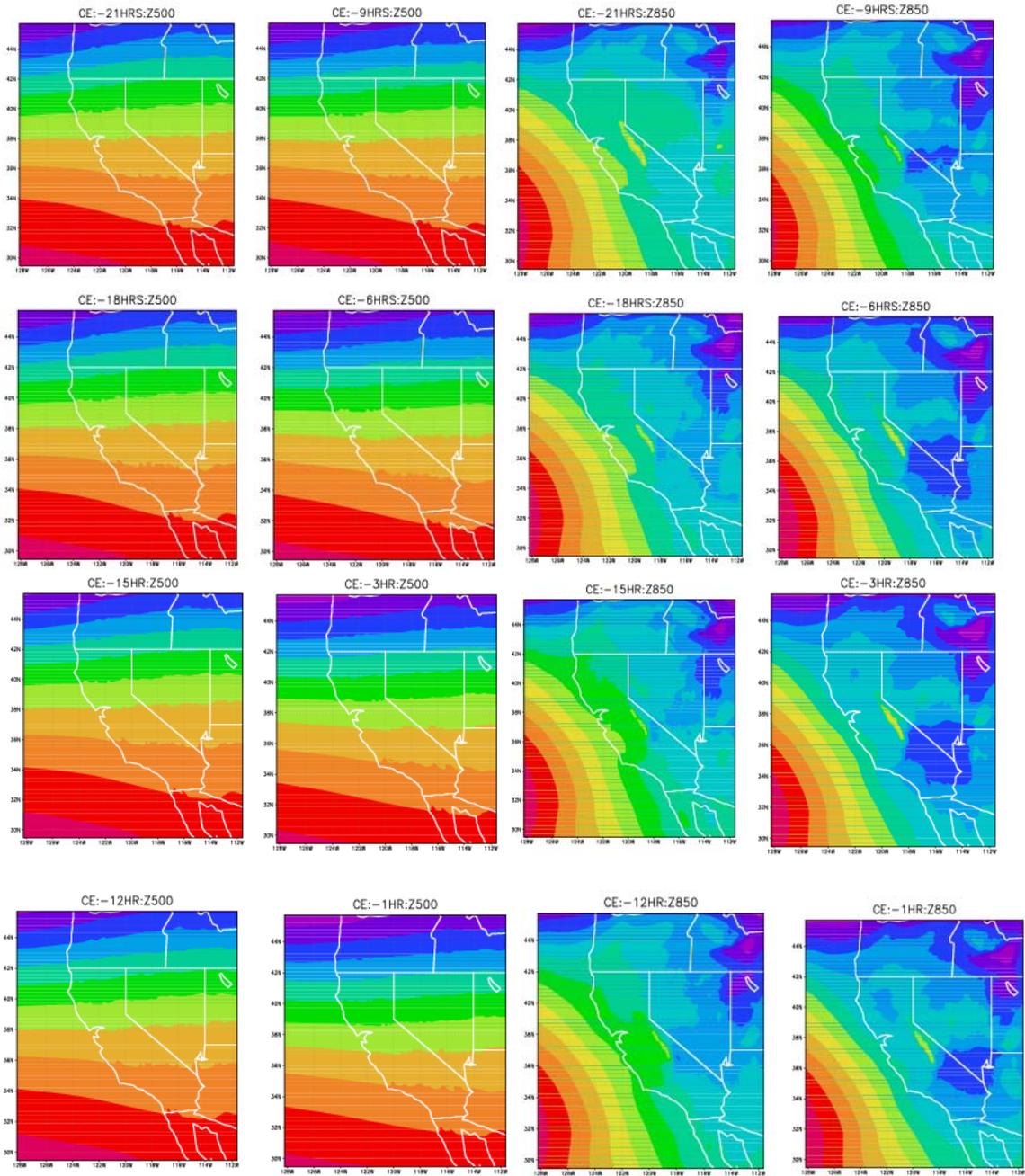
Figure 15. 500 hPa and 850 hPa geopotential height fields at 21, 18, 15, 12, 9, 6, 3, and 1 hours before the detection of CE event
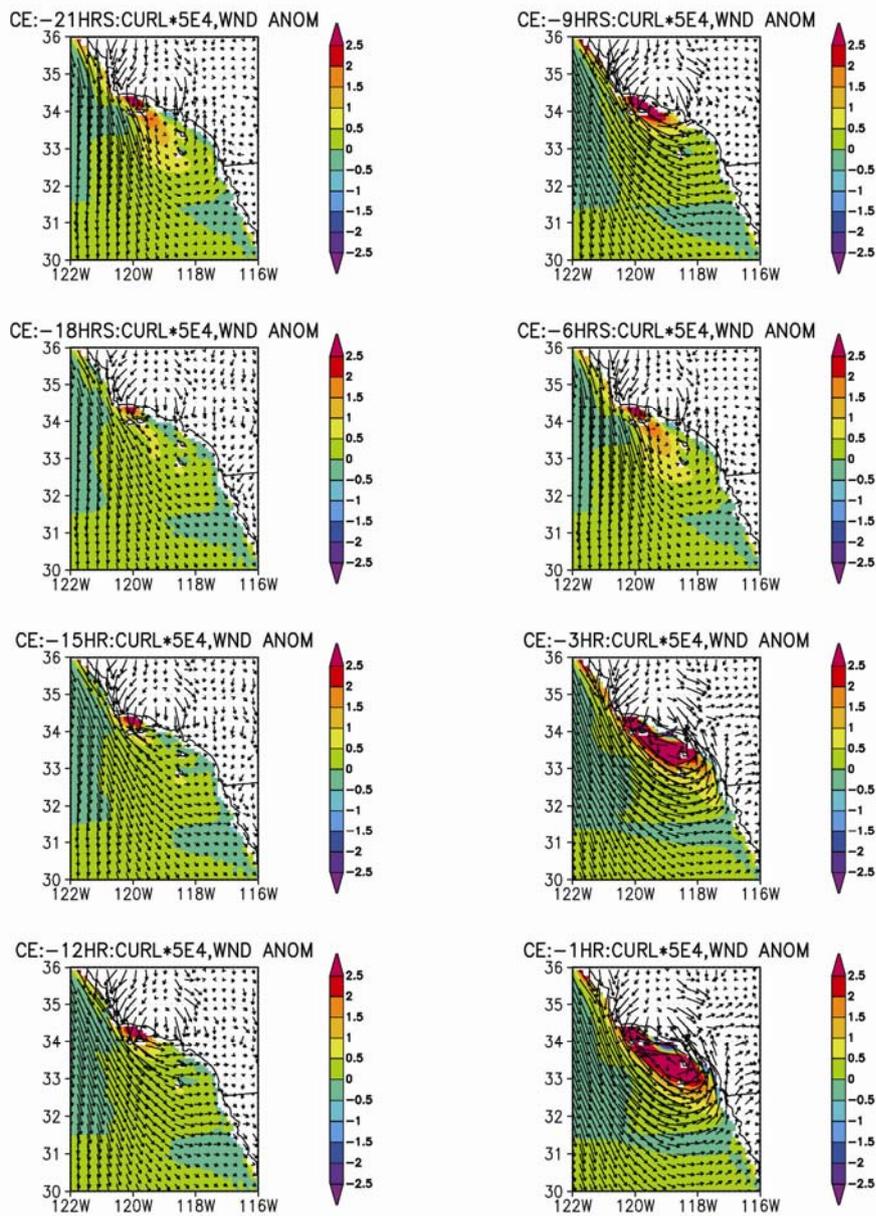
Figure 16. Anomaly wind vector and vorticity at 21, 18, 15, 12, 9, 6, 3, and 1 hours before the detection of CE event
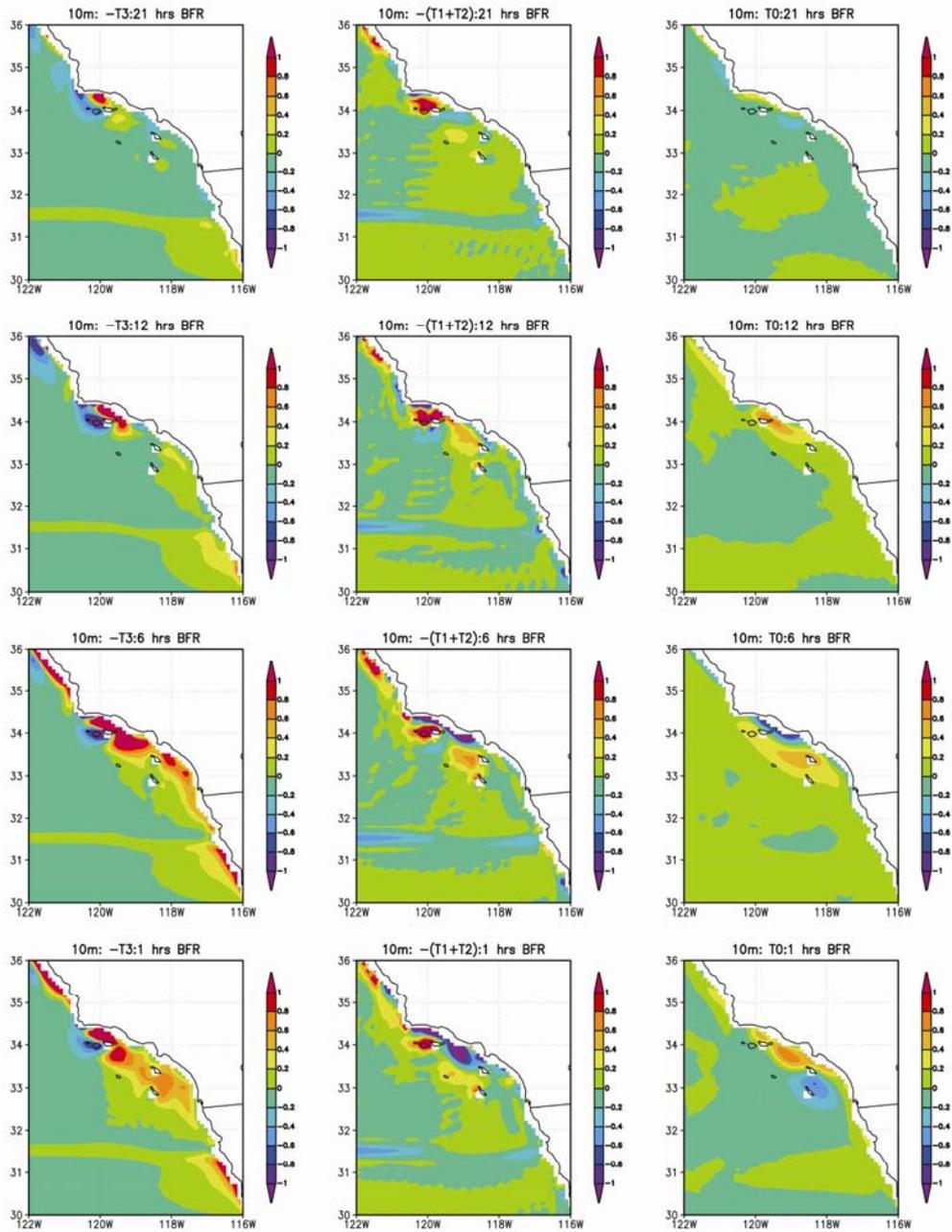
Figure 17. Vorticity budget at 10m above ground at 21, 12, 6 and 1 hour prior to the recognition phase. T3 is a stretching term, T1+T2 is the horizontal advection term and T0 is a local time change or the storage term.
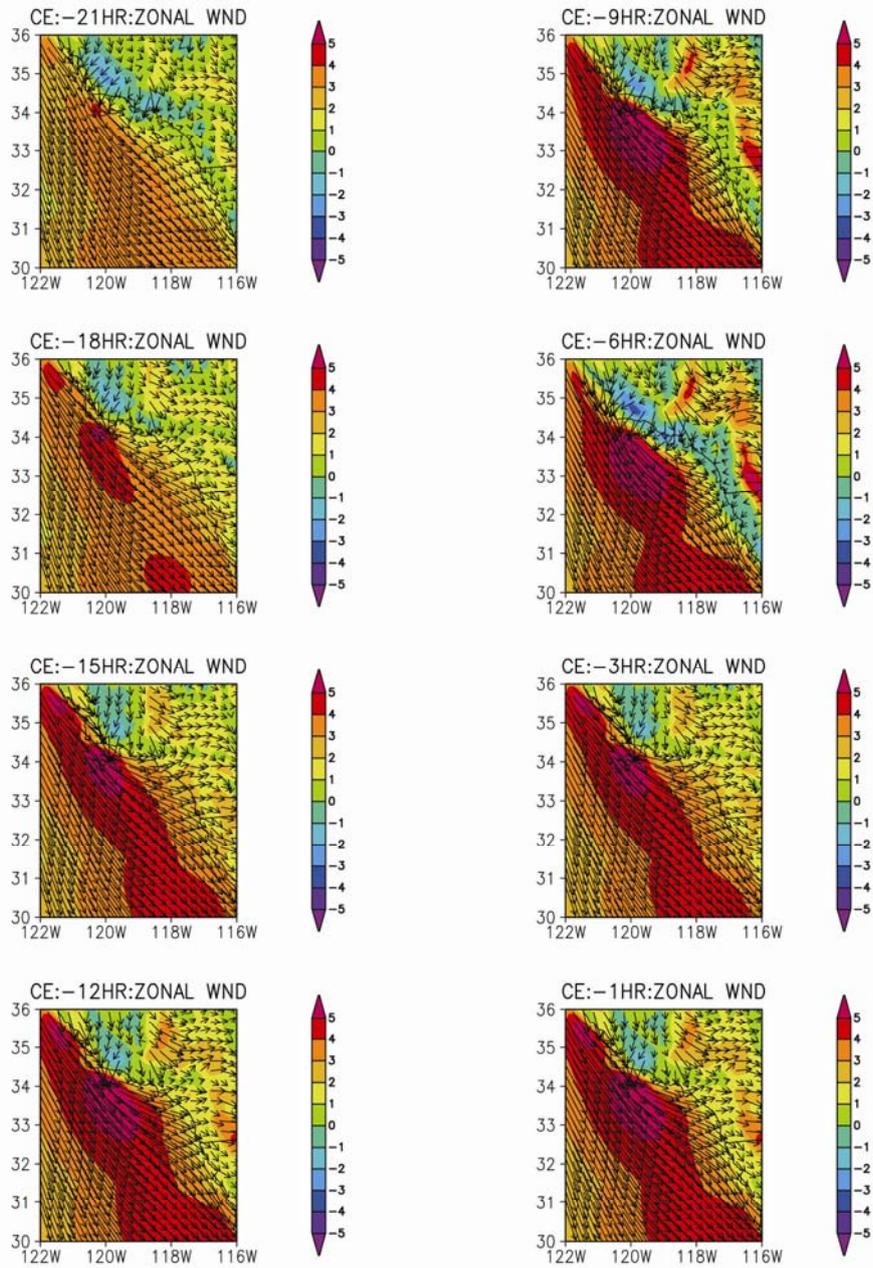
Figure 18. Near surface wind vector at initial stages of the eddy formation. Shading is for the east-west component of the wind highlighting the development of the off shore wind (cold colors).
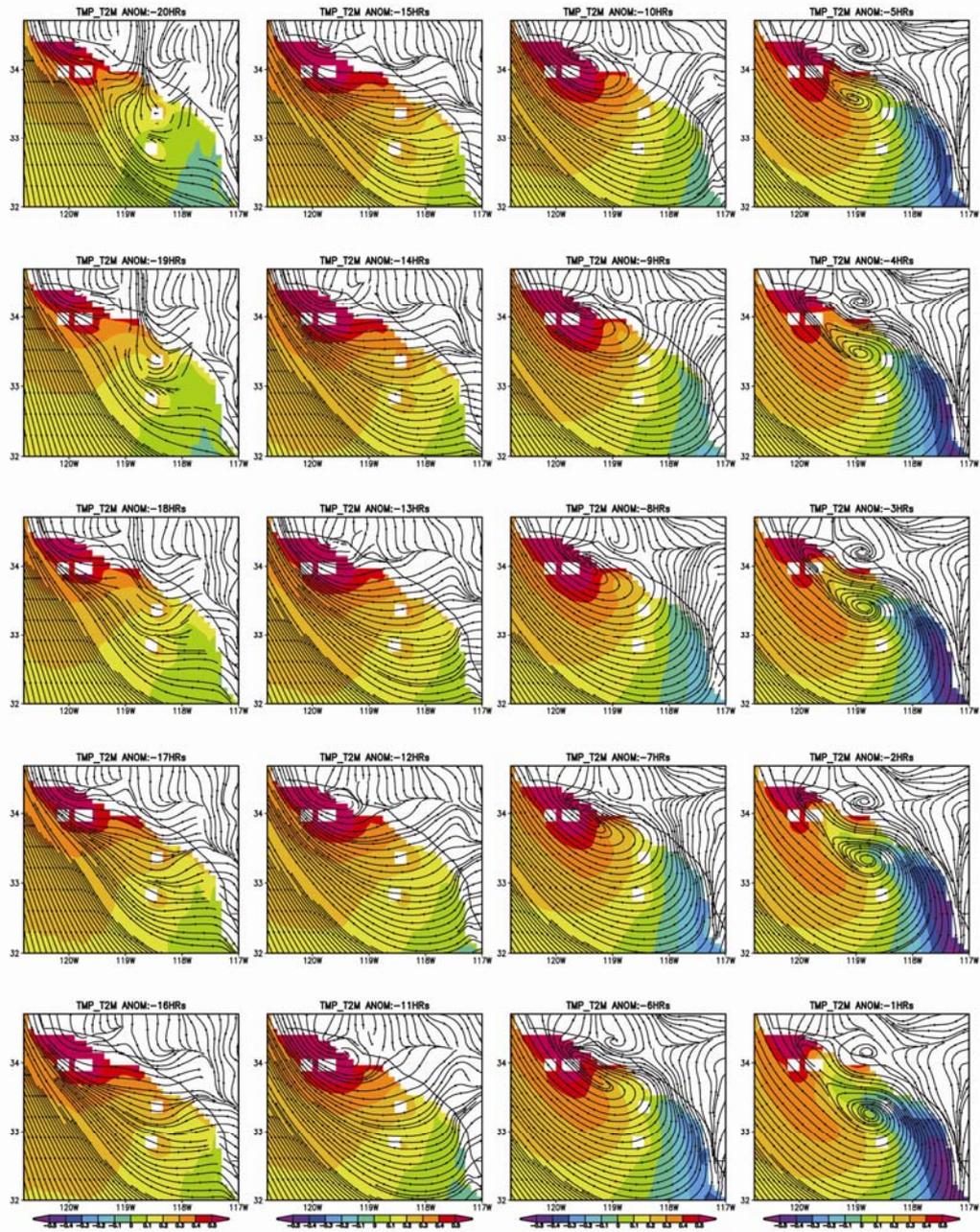
Figure 19. Composite evolution of near surface air temperature anomaly during the initial phase.
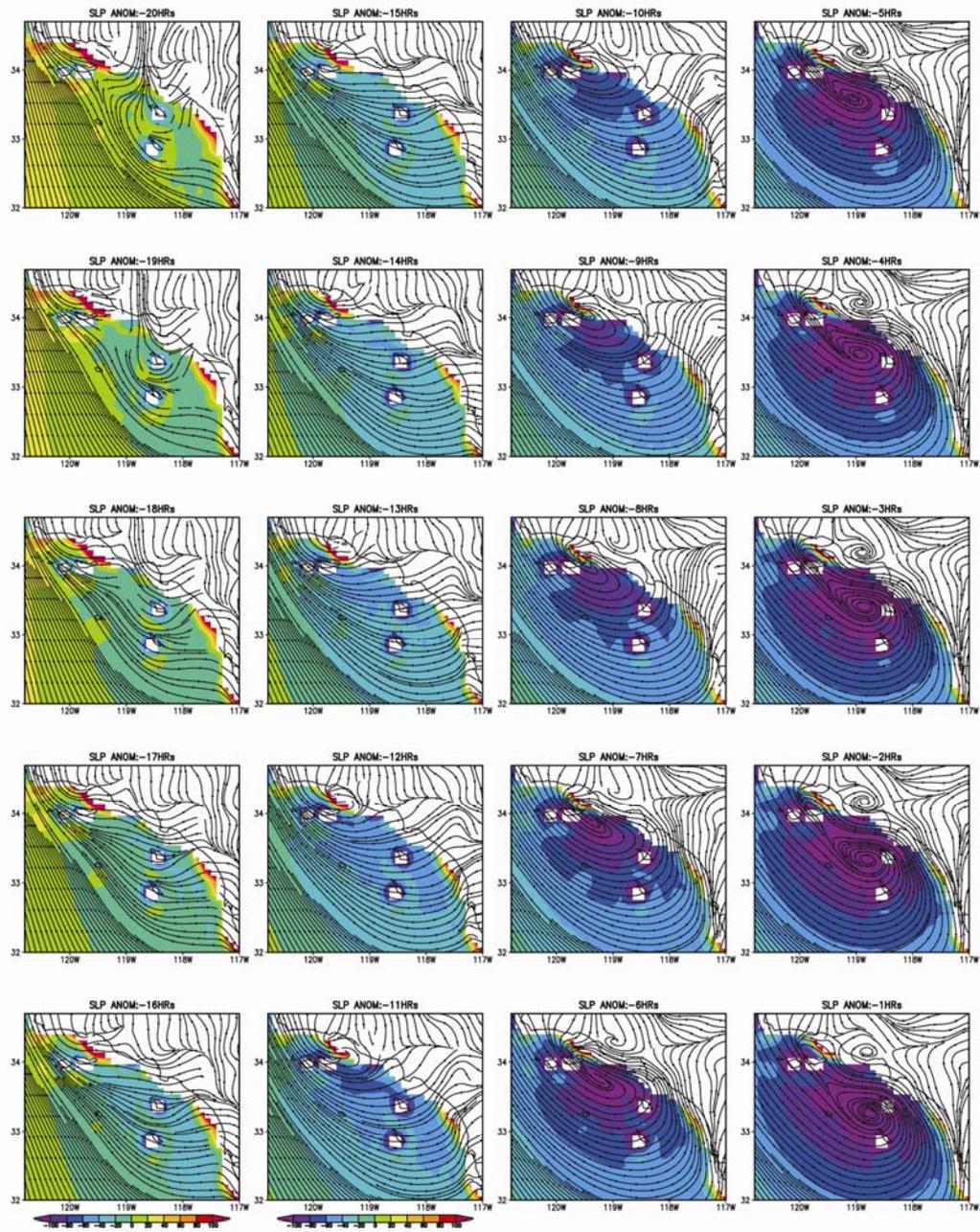
Figure 20. Hourly variation of surface pressure anomaly during the initiation phase of the Catalina eddy obtained from composite.
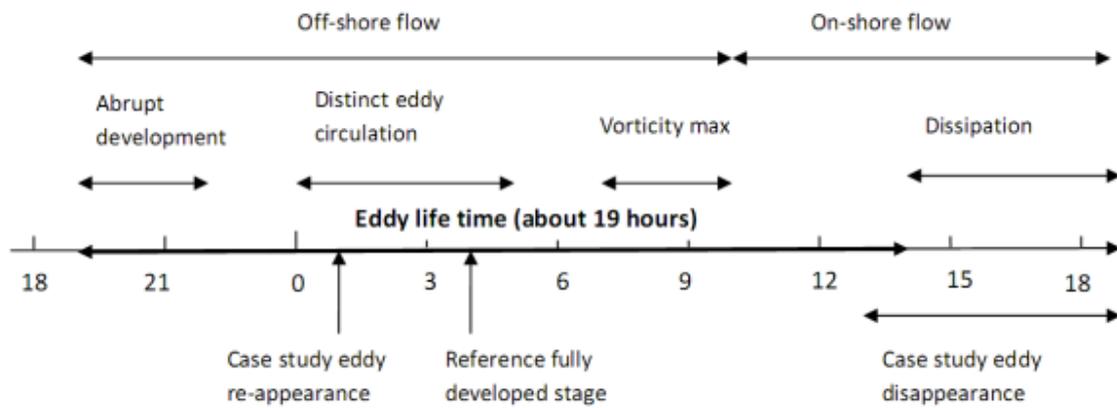
Figure 21. Illustration of the Catalina Eddy life cycle. Horizontal axis is time in local Pacific Standard Time.
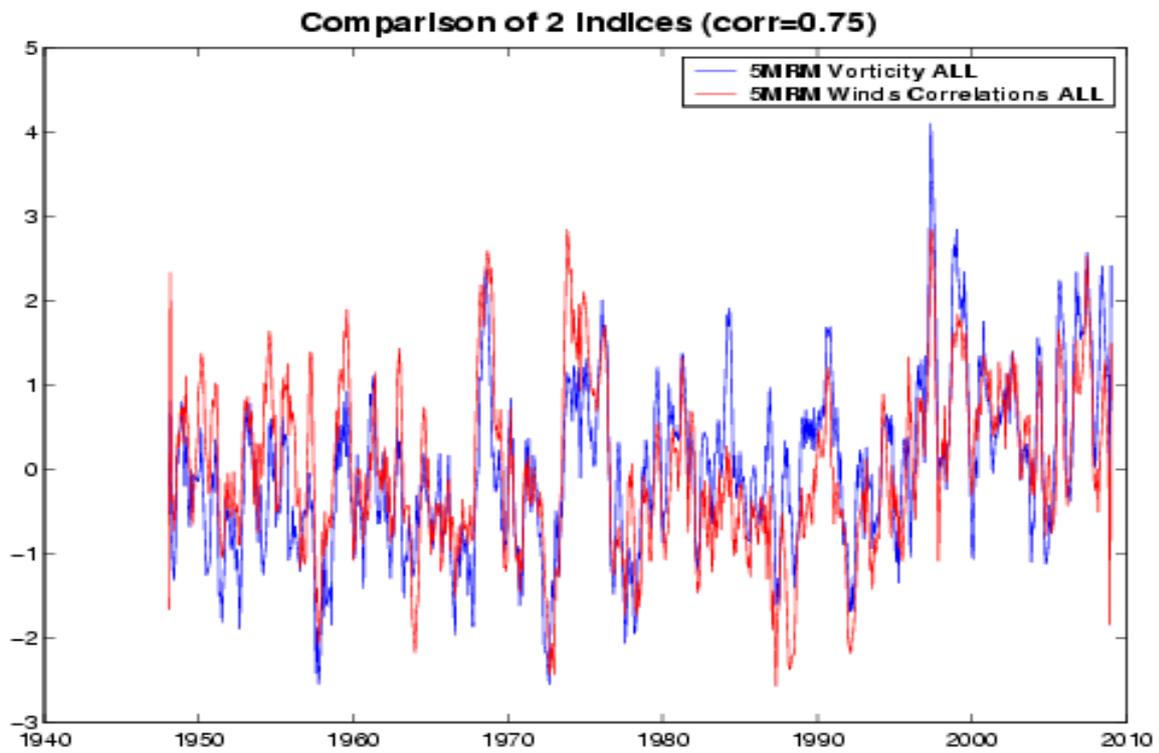
Figure 22. Comparison of two CE indices; I2.1 Monthly averages of Vorticity averaged over CE region and I2.2 Monthly averages of Winds Correlations averaged over CE region. The correlation between the two indices was 0.75.
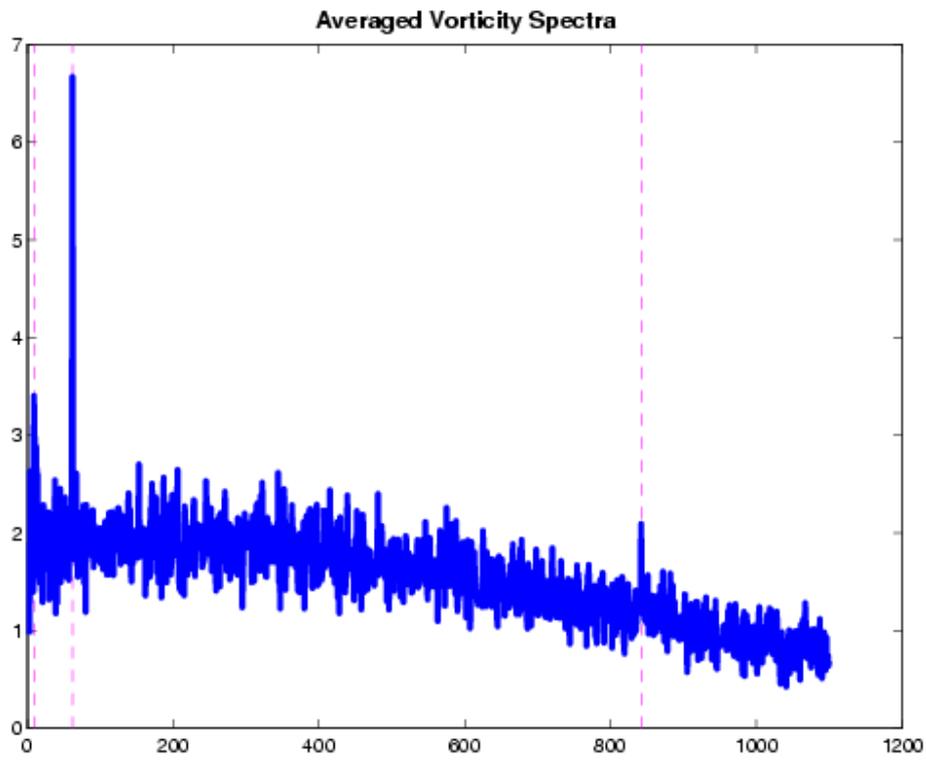
Figure 23. Temporal spectra of vorticity averaged over the Southern California Bight region (32°N – 34°N, 121° W- 117° W).

Regression of CE index onto R-2 monthly Z500 (1970–2008)
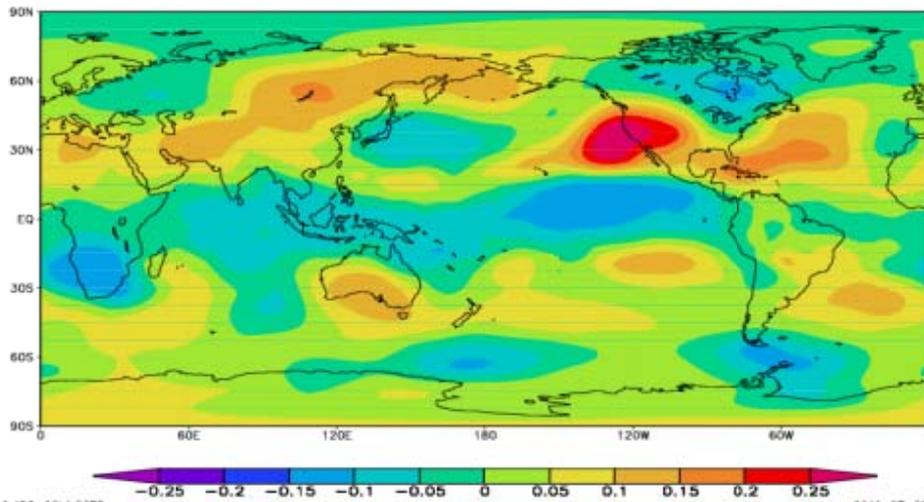
**Regression of CE index onto R-2 monthly T2m (1970–2008)**
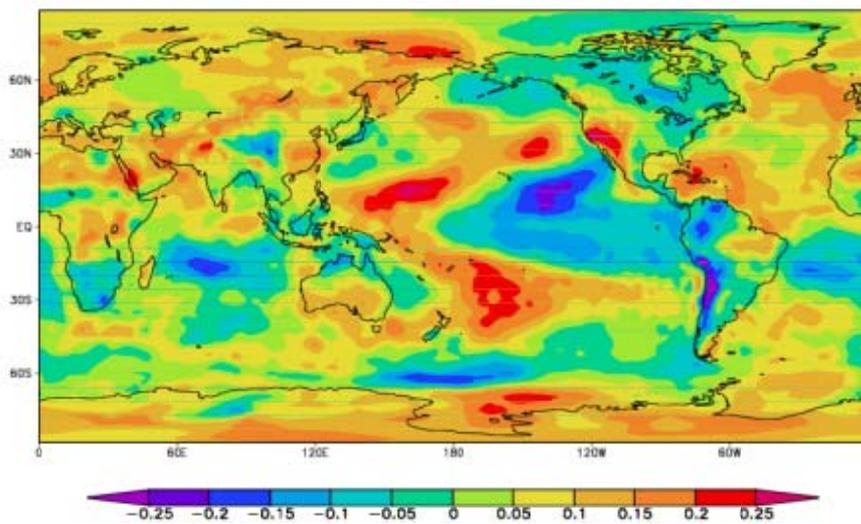
Figure 24. Regression of CE index onto NCEP/NCAR reanalysis monthly mean 500 hPa height (top) and near surface temperature (bottom).